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AN INVESTIGATION OF SEISMIC WAVE PROPAGATION IN THE EASTERN UNITED STATES

July 1968

Prepared For

Geophysics Division

Air Force Office of Scientific Research

Arlington, Virginia 22209

By

David E. Willis

Geophysics Laboratory

Willow Run Laboratories

Institute of Science and Technology

The University of Michigan

Ann Arbor, Michigan

Sponsored By

Advanced Research Projects Agency

Nuclear Test Detection Office

Project VELA UNIFORM

ARPA Order No. 292, Amendments 32 and 37

NOTICES

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Preface

The research reported in this paper was conducted at the Geophysics Laboratory of the Institute of Science and Technology, The University of Michigan over a six year period. Many members of the Geophysics Laboratory assisted in the field measurement and data analysis program. The author gratefully acknowledges their assistance and the vital role they played in this research.

The author would also like to thank Dr. James T. Wilson and Dr. John M. DeNoyer for their valuable contributions during the course of this research. This report was prepared as a partial fulfillment of the requirements for a doctoral degree at The University of Michigan.

This research was supported by the Advanced Research Projects Agency as a part of Project VELA UNIFORM and was monitored by the Air Force Office of Scientific Research under Contracts AF49(638)-1170 and AF49(638)-1759.

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Abstract

This paper describes the travel-time anomalies and attenuation losses of seismic compressional waves generated by a series of underwater explosions in the eastern United States. The efficient tamping of the shots fired in water provided a seismic source that could be detected at much larger ranges than could be accomplished by equivalent sized shots fired underground.

A number of mobile field recording stations equipped with three-component matched short period seismometers and magnetic tape recorders were used to record 273 individual shots. A total of 1,295 recordings were obtained along a reversed profile extending from International Falls, Minnesota to the Atlantic Coast. The underwater shots were fired in Lake Superior and the Atlantic Ocean. Precise travel times were obtained by recording radio time signals at all of the recording stations.

An analysis of the travel-times of the seismic waves disclosed that the earth's crust varies in thickness from 28.1 km near the Atlantic Coast in North Carolina to 50.4 km near the Keweenaw Peninsula in upper Michigan. The crust in North Carolina was found to be comprised of a single layer with a compressional wave velocity of 6.0 km/sec. A two-layer crust with compressional wave velocities of 6.2 and 7.7 km/sec was disclosed in upper Michigan. Travel time residuals across the Appalachian Mts. would indicate a mountain root system similar to that found under the Rocky Mts.

The Lake Superior area was found to be more efficient in the coupling of energy into seismic waves by the underwater shots than in the Atlantic Ocean. This is believed due mainly to the bubble pulse phenomenon and the reinforcement caused by reverberation of sound waves between the lake bottom and the water's surface.

A new technique of displaying seismic attenuation data in three dimensional space was developed that permits presentation of large quantities of data in a concise manner, hence aiding in the interpretation of the data. Consistent anomalies in the spectrum of the seismic waves were found that could be correlated with various parameters in the source region and with propagation effects.

The detailed spectral studies that were made as a function of distance from the source disclosed that the attenuation of the seismic waves could be expressed by the equation

$$A = A_o X^{-n} e^{-\alpha f X}$$

where A = amplitude, X = distance, n = geometrical spreading factor, α = absorption coefficient and f = frequency. The crustal attenuation data in the Lake Superior area were found to have a value of n = 1.5 and $\alpha = 2\text{-}3 \times 10^{-3}$. Upper mantle compressional wave attenuation data gave a value of n = 1 and $\alpha = 2 \times 10^{-4}$.

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Introduction

The attenuation or decrease in amplitude of seismic waves with increasing distance from the source has been a problem of considerable interest to seismologists and other scientists for many years. Predicting damage due to quarry shots, areas affected by earthquakes and the general problem of detecting and locating underground nuclear detonations are just a few areas where this subject has specific applications. Many factors can contribute to the seismic loss of energy with distance. An analysis of the precise travel times of seismic waves allows seismologists to compute the structure of the earth's interior and to determine the seismic velocities of the structural components. Information concerning the latter provides insight into the composition of the materials that comprise these features. For a complex problem of this nature sound theoretical hypotheses and detailed field measurements are necessary before one can ascertain the significant contributing factors. It is the purpose of this study to present the results of an extensive investigation on the propagation of seismic waves in the eastern United States, to show how these waves are attenuated and to determine the major crustal and upper mantle structure.

Discussion of Wave Types

Factors that contribute to the attenuation of seismic waves are listed below:

1. absorption (i.e., conversion of mechanical energy into heat)
2. scattering
3. geometrical spreading
4. reflection and refraction (including conversion of wave types)

Absorption of seismic waves should theoretically be of a solid friction or viscous nature or a combination of both. If solid friction is the significant factor, an exponential amplitude decay proportional to the first power of the frequency ($A \approx e^{-\alpha f x}$ where α = absorption coefficient f = frequency, and x = distance) would be expected (Howell 1963). Viscous type friction on the other hand would result in an exponential decay proportional to the second power of the frequency. Born [1941] has shown for low frequency data (below 150 cps) that viscous absorption is unimportant and if scattering occurs from small inhomogeneities less than a wavelength in dimension that the loss is of the same nature as viscous absorption, e.g., $A \approx e^{-\alpha f^2 x}$. Selective scattering due to reflections and refractions on prominent irregularities would have more pronounced effects on wavelengths of the same order as the irregularities. Multiple wave paths may cause constructive or destructive interference resulting in scatter of the attenuation data.

For body waves (compressional and shear) geometrical spreading would cause a decrease in amplitude that would be proportional to the inverse first power of the distance ($A \approx 1/x$) between source and receiver. Surface waves

(Rayleigh and Love) attenuate like the inverse one-half power of the distance. Head wave amplitudes of compressional waves (see Heelan [1953], Werth [1962] or Donato [1964]) should fall off like $(x')^{-2}$ where $x' = (xL^3)^{1/4}$ with x equal to the horizontal distance from the source to the recording station and L the distance traveled in the refractor. Hence for large distances head waves should attenuate approximately like x^{-2}. The last major factor contributing to the attenuation of seismic waves is the energy loss that a wave encounters upon reflection and refraction (including conversion of wave type) at a velocity discontinuity. This factor has been described in some detail by the classical works of Knott [1899], Macelwane [1936], Zoeppritz [1919], and others.

This study will be restricted primarily to the attenuation of body waves of the compressional-dilatational type. Travel time data will be included to provide as much correlation with geological parameters as possible. The following section contains a description of the field data collected over a five year time period that was utilized in this investigation.

Field Measurement Program

A reversed long-range seismic refraction profile was conducted along a northwest-southeast line approximately 2000 km long in the eastern United States. The line extended from International Falls, Minnesota to the North Carolina

Atlantic Coast. This exercise comprised a series of separate experiments with many participants. The data collected by University of Michigan field crews is unique in that it is one of the longest reversed profiles with close station control in existence, and it is also the only reversed profile over the Appalachian Mountains. The following series of shots comprise this study: (see Table I).

Details of these experiments are included in Appendix A. The University of Michigan operated from two to nine portable seismograph stations to record these shots. Each station was equipped with FM magnetic tape recorders, matched short-period three-component seismometers and short wave radios to record radio time signals. Details and specifications of this equipment are discussed in Appendix B. The general technique employed to obtain the detailed close station control was to assign mobile stations to a given segment along the main refraction profile. The shots were usually fired at night ranging over a time period from several days to a month. The recording stations were moved each day so as to cover as much ground within their segment as possible. Sites were located on bedrock wherever available and in areas to avoid cultural activity. At the larger distances with low-signal levels the site selection was very critical since poor signal-to-noise levels made it very difficult to detect the first compressional wave arrivals.

Table I. Explosion Seismic Experiments

Experiment	Field Stations	Recording Sites	Shots	Recordings	Range Covered (km)
1962 North Carolina	3	4	20	37	27-248
Lake Superior, 1963	7	31	78	473	16-600
Lake Superior, July 1964	3	8	32	96	30-200
Lake Superior, Oct 1964	7	21	10	52	72-762
ECOOE	9	30	92	315	70-1600
CHASE III	9	9	1	3	360-1490
CHASE IV	6	6	1	6	971-1980
CHASE VII	8	8	1	6	512-1342
Early Rise	8	121	38	308	44-1681

Travel Times and Crustal Structure

The shot point and recording station locations are shown in Figure 1 where data reported in this paper were obtained. The series of shots fired in the Atlantic Ocean include the CHASE series, the 1962 North Carolina Experiment, and the ECOOE project. The Lake Superior shots mentioned in the previous section are also shown in this figure.

Reduced travel time graphs were prepared for the first compressional wave arrivals recorded at the stations shown in this map. These are presented in Figures 2 through 12. Graphs of this type allow one to examine more carefully the deviations or anomalies in travel times than would be possible to see in the standard time vs distance graphs.

Figure 2 is a composite made up of first compressional wave arrivals recorded at Jacksonville, Potters Hill, and Dudley, North Carolina for the shots fired during the 1962 North Carolina experiment. The University of Michigan station locations and first arrival times may be found in a paper by Meyers, et al. [1965]. Using a crustal velocity of 6.0 km/sec for the reduced travel time curves it can be seen that velocities of 5.8 and 7.9 km/sec for the crustal layer and upper mantle velocity gives a reasonable fit to the data. Figures 3 and 4 show the reduced travel time residuals for the individual stations of the ECOOE Southern Profile shots. The station locations and first arrival times for the 150, 250, and 400 km University of Michigan stations on the Southern Profile and the 150, 300, and 325

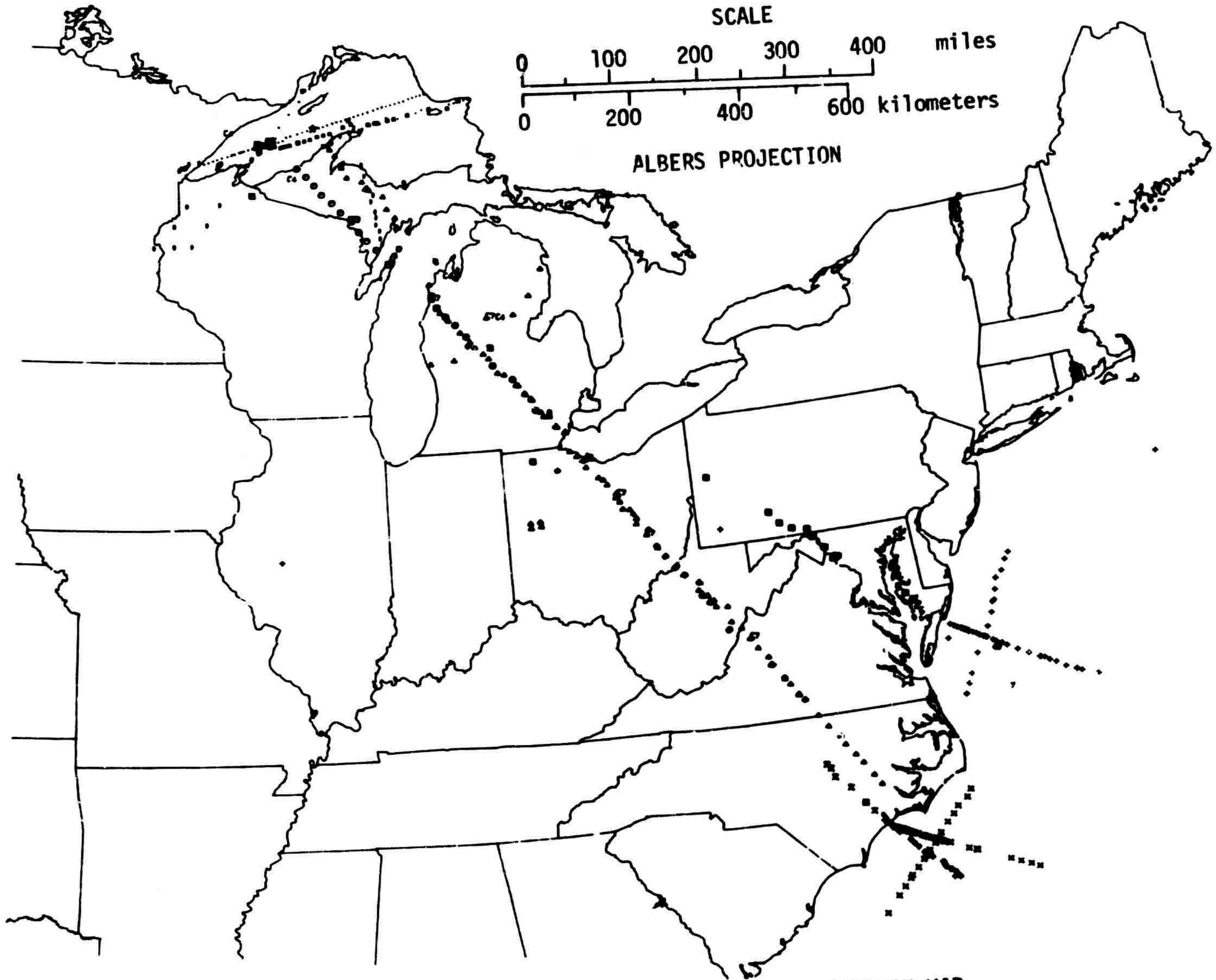

FIGURE 1. SHOT POINT AND RECORDING STATION LOCATION MAP.

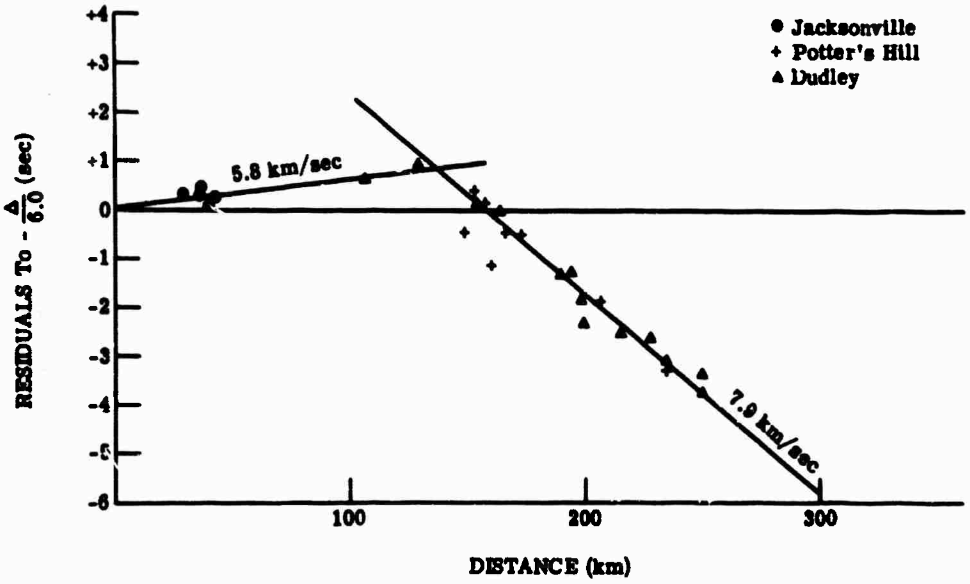

FIGURE 2. TRAVEL-TIME RESIDUALS: 1962 NORTH CAROLINA EXPERIMENT.

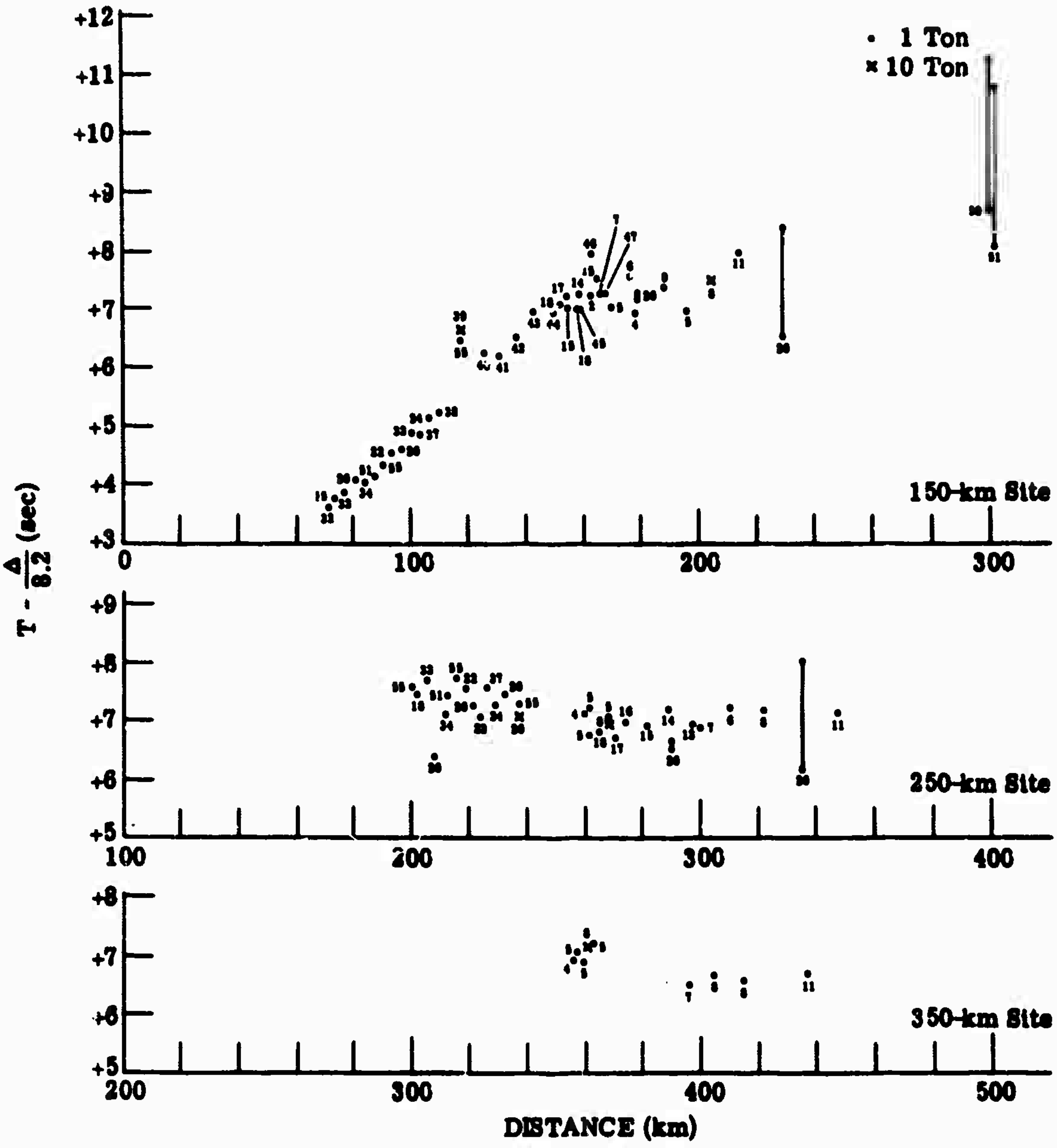

FIGURE 3. TRAVEL-TIME RESIDUALS: SOUTHERN PROFILE, 150-, 250-, AND 350-km SITES

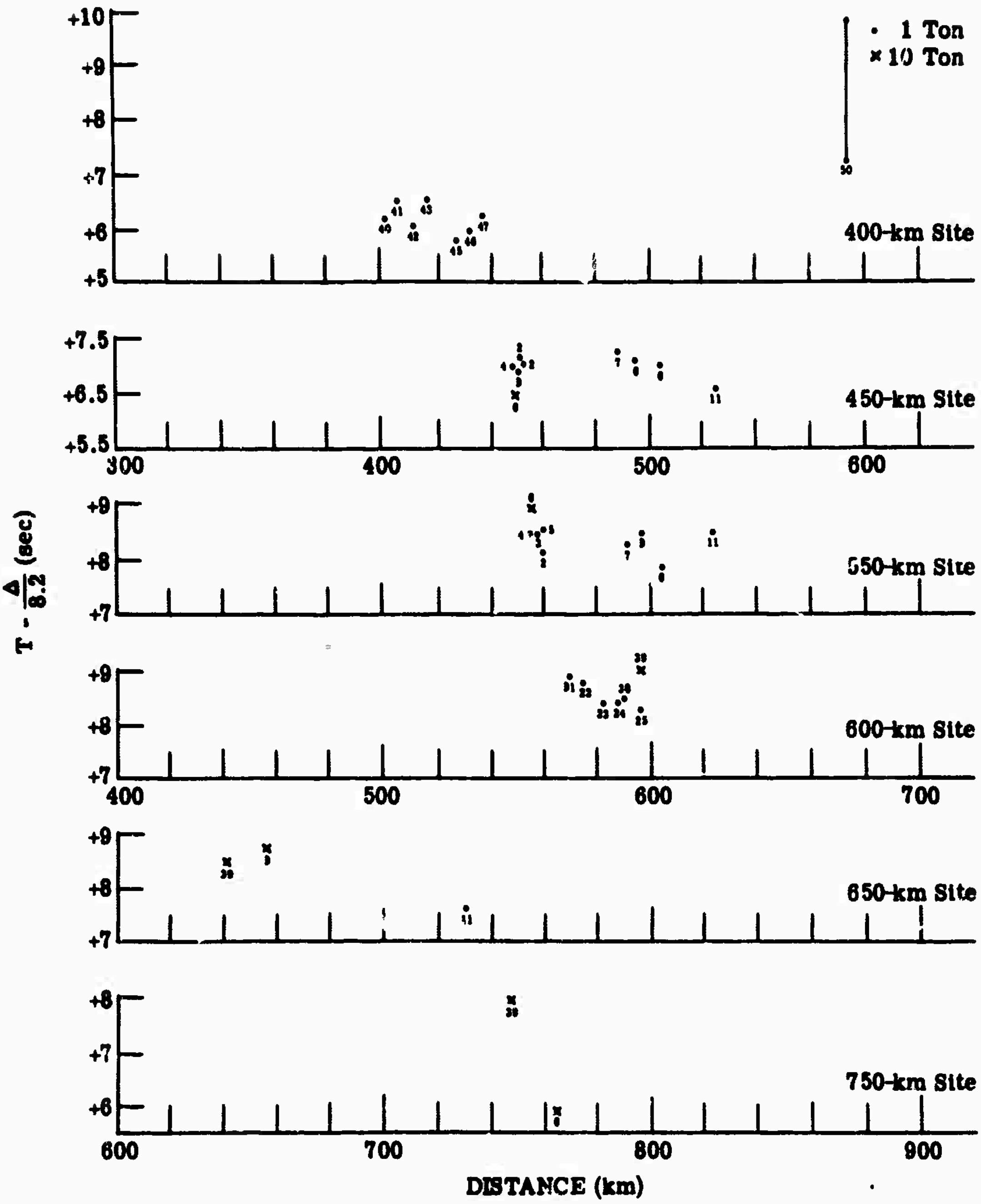

FIGURE 4. TRAVEL-TIME RESIDUALS: SOUTHERN PROFILE, 400-, 450-, 550-, 600-, 650-, AND 750-km SITES.

km stations on the Northern Profiles of the ECOOE series may be found in a paper by Hales, et al. [1967]. This information was supplied by the University as a part of their contribution to the cooperative experiment conducted by the North American Explosion Seismology Group. Figure 5 is a composite of the data for Figures 2 through 4 and include the CHASE III and VII shots. An upper mantle velocity of 8.2 km/sec was used to compute the data shown in the latter figures.

The data shown in Figure 5 disclose three branches in the travel time curve. These may be fit reasonably well to the following:

$$T_1 = \Delta/2.0 + 0 \quad (0\text{-}30\text{ km}) \quad h_1 = 0.49\text{ km (assumed)} \tag{1}$$

$$T_2 = \Delta/6.0 + 0.48 \quad (30\text{-}150\text{ km}) \quad h_2 = 28.69 \tag{2}$$

$$T_3 = \Delta/8.2 + 7.0 \quad (150\text{-}1000\text{ km}) \tag{3}$$

where T = travel time of first compressional wave arrival

Δ = distance between shot point and recording station

h = layer thickness.

The velocity for the first branch or upper layer is assumed since we have no data at close ranges. A velocity of 2.0 km/sec is quite reasonable for a surface layer. An error in this assumption would produce no serious discrepancies for the total thickness of the crust since the intercept of the second branch (0.48 sec) is small. These data agree quite well with those reported by Hales which will be discussed later.

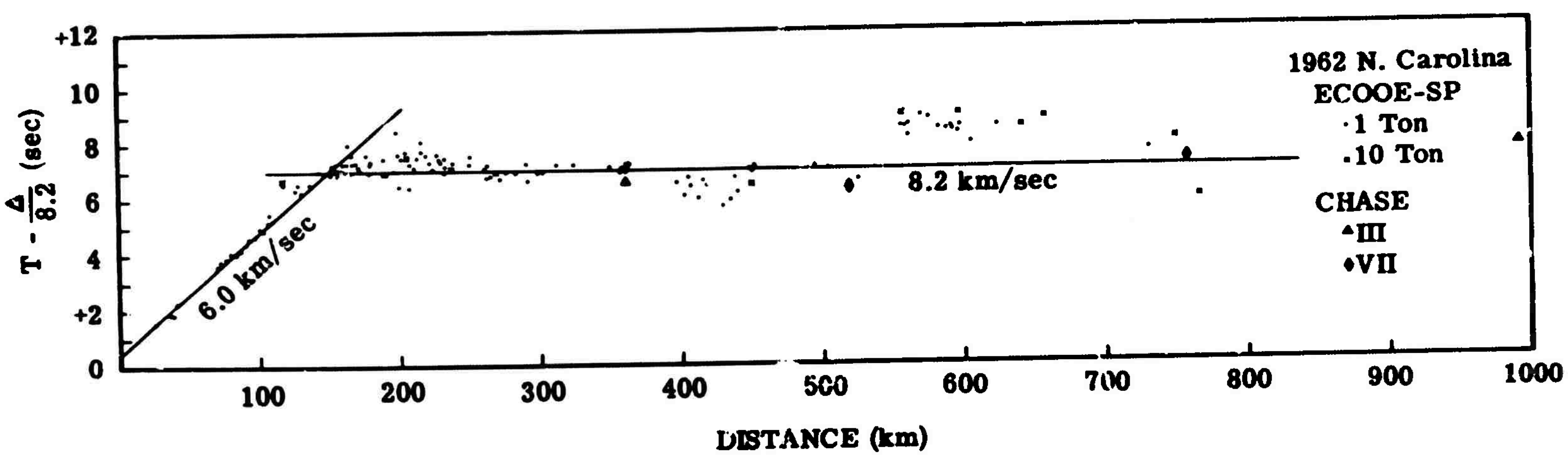

FIGURE 5. COMPOSITE TRAVEL-TIME RESIDUALS FOR 1962 NORTH CAROLINA EXPERIMENT, OOE SOUTHERN PROFILE AND CHASE III AND VII.

Considerable scatter is evident on the upper mantle (8.2 km/sec) branch of the travel time curve shown in Figure 5. However, when observing the individual station residuals in Figures 3 and 4 a systematic deviation can be seen. The 150, 250, and 350 km sites have a residual of approximately 7 to 7.5 seconds. The latter two sites indicate a slightly higher P_n velocity. The 400 km site which is located in the Piedmont Belt shows a significantly lower residual of approximately 6 seconds. This could be accounted for by a thinner crustal section related to the Piedmont Triassic Belt. The western edge of the Appalachian root lies approximately halfway between the 450 and 550 km sites. An increase in residuals from 7.5 to 8.5 seconds seen at these two sites could be caused by the effect of the root. This would also effect the next two sites (600 and 650 km) which would fall in the shadow of the root. The stations at greater distances should not show any anomalous effects of the root since the ray paths would be below the projection of the root system which is reported by James [1966] to reach 60 km. As can be seen in Figure 5 the residuals beyond 700 km average about 7.5 seconds.

Similar travel time residuals for the ECOOE Northern Profile are shown in Figures 6 through 8. Unfortunately the shot point location accuracy of the shots fired off the continental shelf are uncertain. The shots are normally located by using water wave arrivals recorded at underwater hydrophones. For the Northern Profile shots the hydrophone buoys

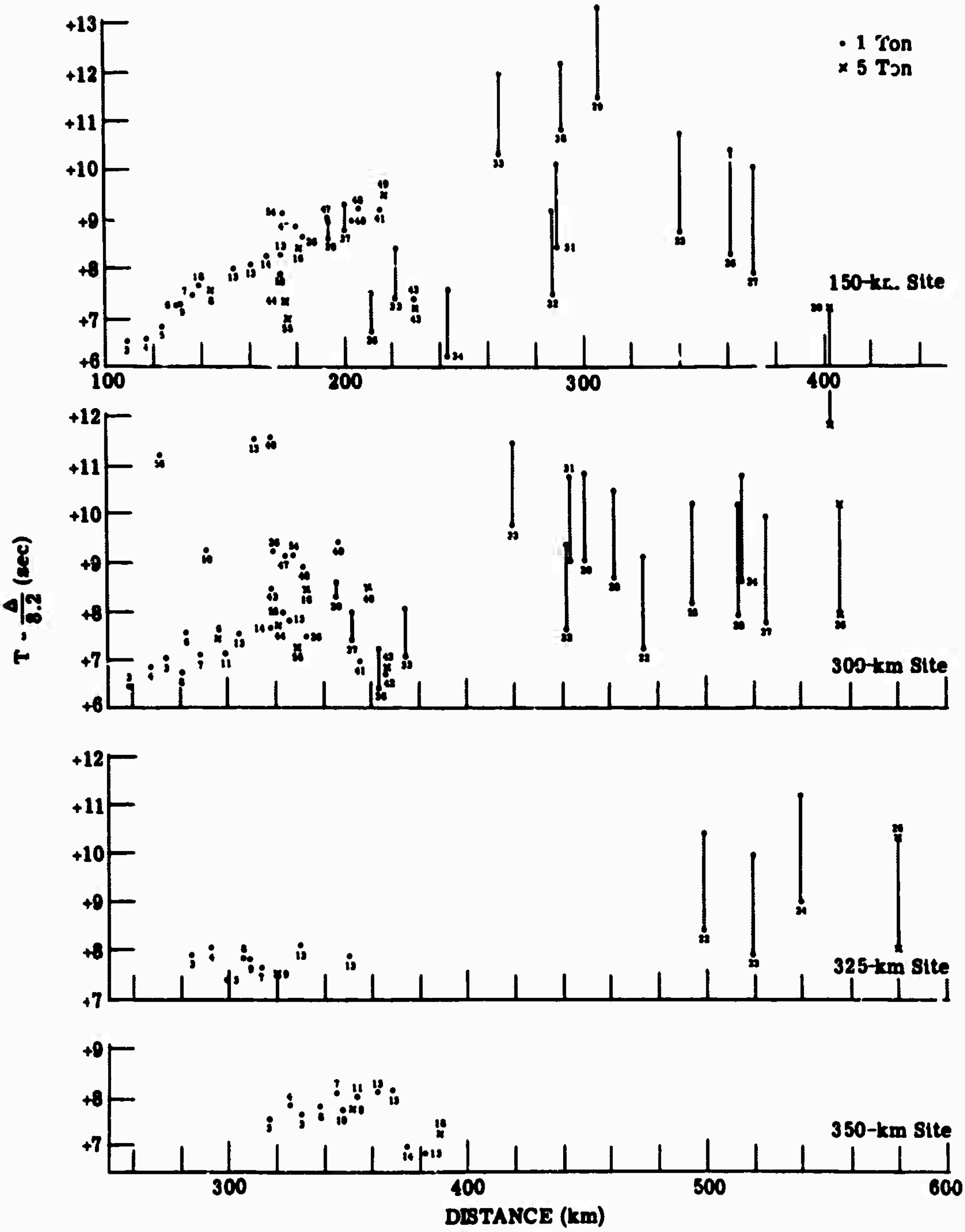

FIGURE 6. TRAVEL-TIME RESIDUALS: NORTHERN PROFILE, 150-, 300-, 325-, AND 350-km SITES

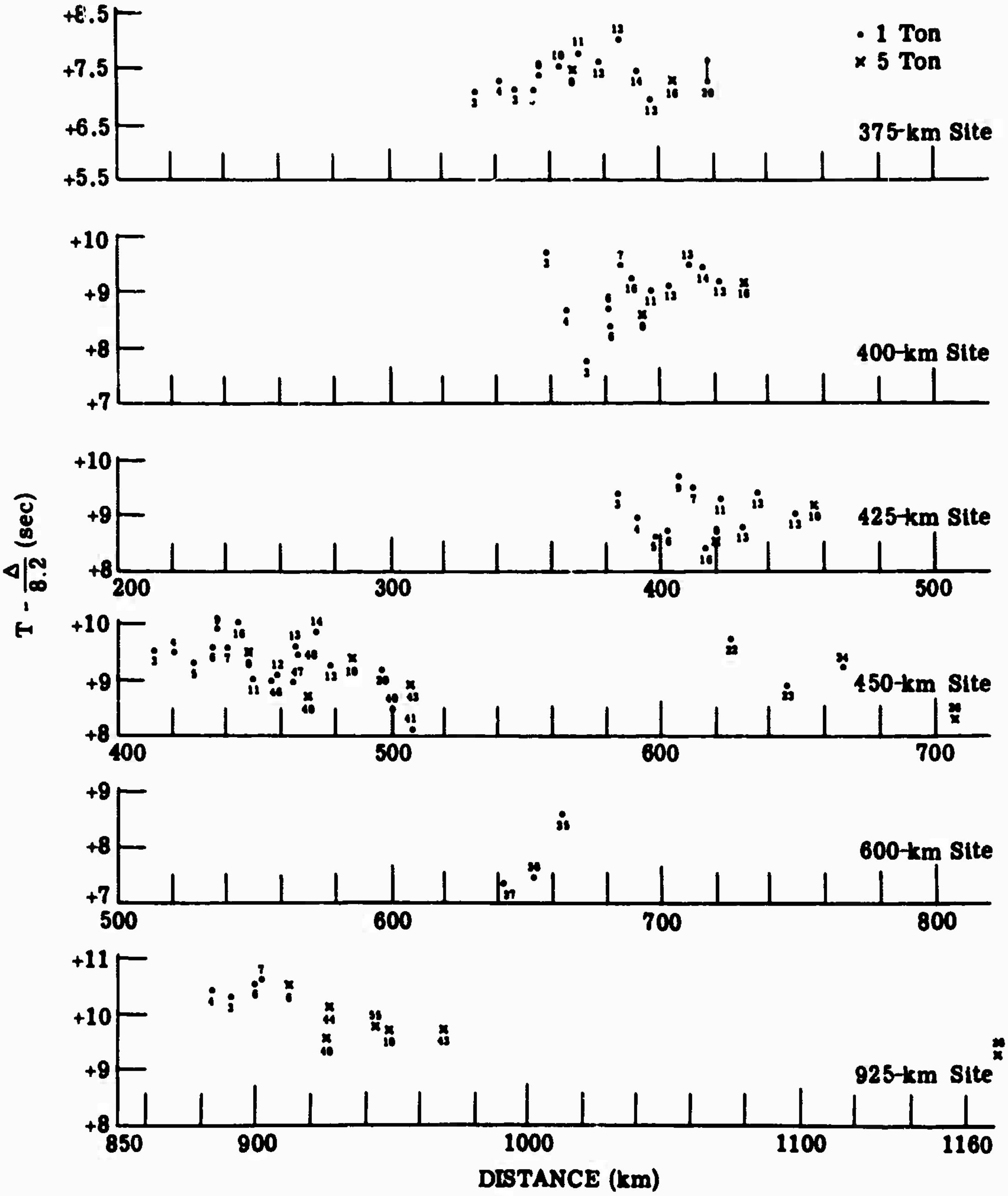

FIGURE 7. TRAVEL-TIME RESIDUALS: NORTHERN PROFILE, 375-, 400-, 425-, 450-, 600-, and 925-km SITES

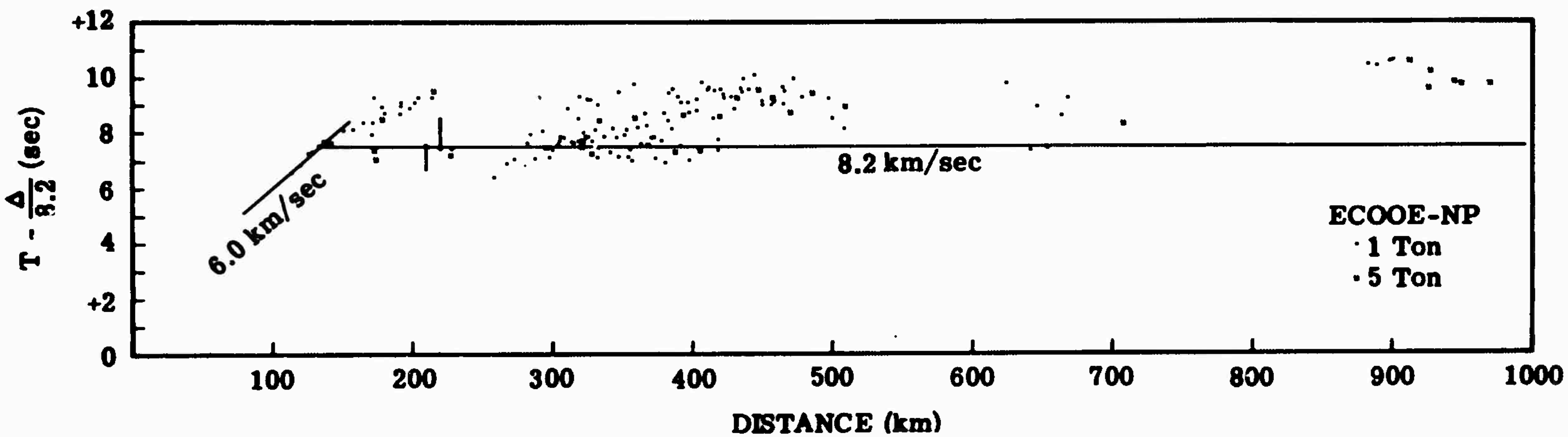

FIGURE 8. COMPOSITE TRAVEL-TIME RESIDUALS FOR ECOOE NORTHERN PROFILE

could n.t be placed in proper locations so that it was necessary to rely on a combination of hydrophone arrays and LORAN C. In analyzing the data afterwards the Southwest Graduate Research Center found that rather large discrepancies existed between locations determined by the two techniques (LORAN C had errors of up to 10% due in part to the fact that part of the transmission path was over land). This uncertainty in shot location can be seen in the range of travel time residuals for shots 20 through 37. Also contributing to the scatter in residuals is the fact that these shots (20 through 37) were fired in deep water (up to 12,240 ft) at a depth of 300 feet while the other shots, located on the continental shelf, were fired on the ocean bottom. All of the data should be corrected to a common datum taking into account the difference in velocities between salt water and the sediments on the continental shelf. Since this particular problem is not germane to the main topic of this report and since the correct velocities to be used in this adjustment are not known precisely, no attempt will be made to correct the travel time residuals for the shots fired off the shelf. The residuals on the P_n branch of the travel time curve for the 150, 300, 325, 350, and 375 km sites average between 7.5 to 8 seconds which would be indicative of a slightly thicker crust than observed on the Southern Profile. The western edge of the Appalachian root is near the 350 km site. The stations at 400, 425, and 450 km show an increase in this residual of about 1 sec

which verifies the shadow effect of the root which was observed on the Southern Profile. No evidence of a thinner crustal section which was observed in the Piedmont section of the Southern Profile was found on this line. However, an examination of the geology of this region shows that the width of the Appalachian Mountain system is considerably narrower in this area and that the Piedmont section is quite thin. Hence, one could not expect a similar seismic anomaly on this profile.

Figure 8 is a composite of all sites. The residuals for shots 22 through 37 have been omitted on this graph. The larger average positive anomaly between 400 and 500 km is quite apparent. However, the large anomaly at 880 to 970 km which was recorded at the Ann Arbor Botanical Gardens Station is rather surprising. A station anomaly of this magnitude was not substantiated by the CHASE IV or VII shots. However, the CHASE III shot recorded at the Botanical Gardens and at a station 160 km from Ann Arbor seem to substantiate it. Two possible explanations can be offered. One is that the events identified as the P_n arrival were incorrectly picked and what was identified as P_n was actually a later larger amplitude P arrival. The other possibility could be a shot point(s) anomaly. The latter does not appear very feasible since CHASE III and IV were fired in the same general area.

The data on the Northern Profile may be divided into three groups and can be expressed mathematically as:

$$T_1 = \Delta/2.0 + 0 \qquad h_1 = 1.68 \tag{4}$$

$$T_2 = \Delta/6.0 + 1.63 \qquad h_2 = 25.84 \tag{5}$$

$$T_3 = \Delta/8.2 + 7.5 \tag{6}$$

All of the participants including the University of Michigan submitted their first arrival times on the ECOOE project to the Southwest Graduate Research Center. Hales [1967] has analyzed these data and reports the following.

<u>Southern Profile (normal to Coast)</u>

$$T = \Delta/1.70 + 0 \qquad h_1 = 0.49 \tag{7}$$

$$T = \Delta/6.03 + 0.55 \qquad h_2 = 30.38 \tag{8}$$

$$T = \Delta/8.13 + 7.32 \tag{9}$$

<u>Northern Profile (normal to Coast)</u>

$$T = \Delta/2.10 + 0 \qquad h_1 = 1.63 \tag{10}$$

$$T = \Delta/5.78 + 1.45 \qquad h_2 = 8.31 \tag{11}$$

$$T = \Delta/6.34 + 2.65 \qquad h_3 = 16.32 \tag{12}$$

$$T = \Delta/7.97 + 6.60 \tag{13}$$

Figure 9 shows the combined residuals for the five- and ten-ton shots of the ECOOE series and all of the CHASE shots. The data shown in this figure and in Figure 5 yield the following.

$$T_1 = \Delta/2.0 + 0 \qquad h_1 = 0.49 \text{ km} \tag{14}$$

$$T_2 = \Delta/6.0 + 0.48 \qquad h_2 = 27.61 \text{ km} \tag{15}$$

$$T_3 = \Delta/8.2 + 7.0 \qquad h_3 = 70.96 \text{ km} \tag{16}$$

$$T_4 = \Delta/8.5 + 11.5 \tag{17}$$

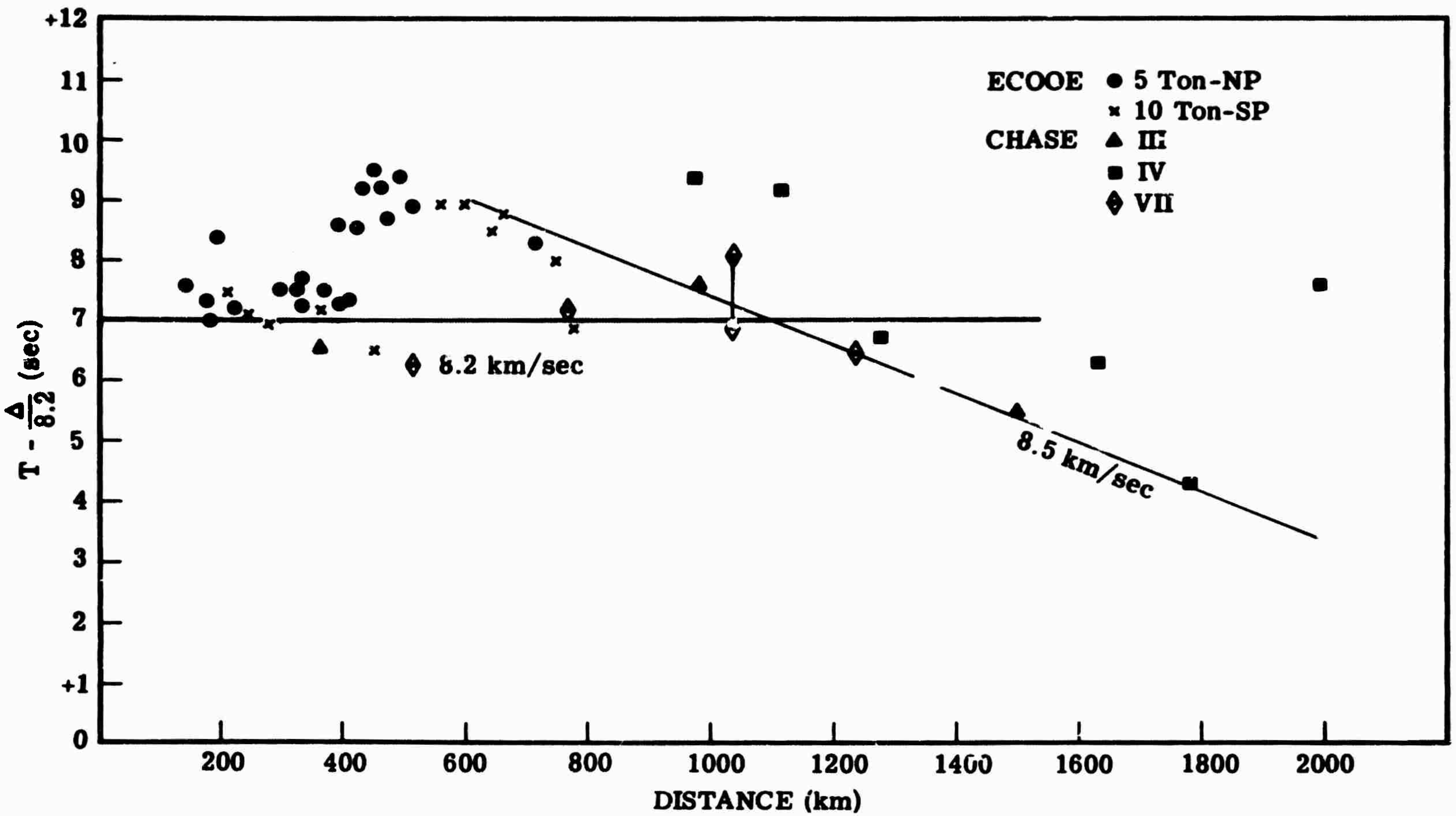

FIGURE 9. COMPOSITE TRAVEL-TIME RESIDUALS FOR 5- AND 10-TON SHOTS OF THE ECOOE SERIES AND CHASE III, IV, AND VII.

A least-squares analyses was made for the travel-time residuals shown for each station in Figures 3, 4, 6, and 7. The deep water shots were omitted because of the large variations in residuals. The resultant velocities and intercepts are shown in Table II. The data for those stations which have only a few recordings or where the distance range is limited should be used with caution.

Table II. ECOOE Compressional Velocities

Site	Velocity (km/sec)	Intercept (sec)
Northern Profile		
150 I	6.8	3.89
150 II	8.2	7.48
150 III	8.0	6.95
300	8.0	6.69
325	8.1	7.42
350	8.7	10.29
375	8.0	6.56
400	7.7	5.97
425	8.2	8.82
450	8.4	10.52
600	Insufficient data	
925	8.5	13.99
Southern Profile		
150 I	6.0	0.57
150 II	8.0	6.81
250	8.5	8.36
350	8.7	9.56
400	8.7	9.20
450	8.2	7.30
550	8.5	10.80
600	9.0	14.69
650	9.1	16.71
750	Insufficient data	

The reversed profile for the series of shots fired in the Atlantic Ocean is made up of a composite of shots fired

in Lake Superior. These shots include the 1963 Lake Superior Experiment, the follow-up sequence of shots in 1964 by The University of Wisconsin, the October 1964 series of shots conducted by the U. S. Geological Survey and the 1966 Project Early Rise shots.

The 1963 Lake Superior Experiment consisted of a series of 38 lake bottom shots fired every 10-15 km along the length of Lake Superior. The University of Michigan's participation in this experiment is described by Willis, et al. [1964], Steinhart [1964], Willis and DeNoyer [1966] and in other papers by the author. Five major profiles were obtained from this experiment only one of which will be utilized in this report. A profile extending southeastward from Houghton on the Keweenaw Peninsula in upper Michigan was obtained by using three mobile stations to occupy 13 recording sites. Shot points located opposite Houghton were recorded at distances ranging from 84 to 258 km. The reduced travel time data from this profile are shown in Figure 10. Some scatter is evident that appears related to individual station corrections. A crustal velocity of 6.4 km/sec was obtained. A lower crustal velocity of 7.5 km/sec was also indicated. However, the scatter of the data made this value somewhat doubtful.

The July 1964 Lake Superior Experiment did not contribute significantly to this study since the recording stations and shots were located to supplement the Eastern and Western Profiles on the 1963 Lake Superior Experiment. The shots

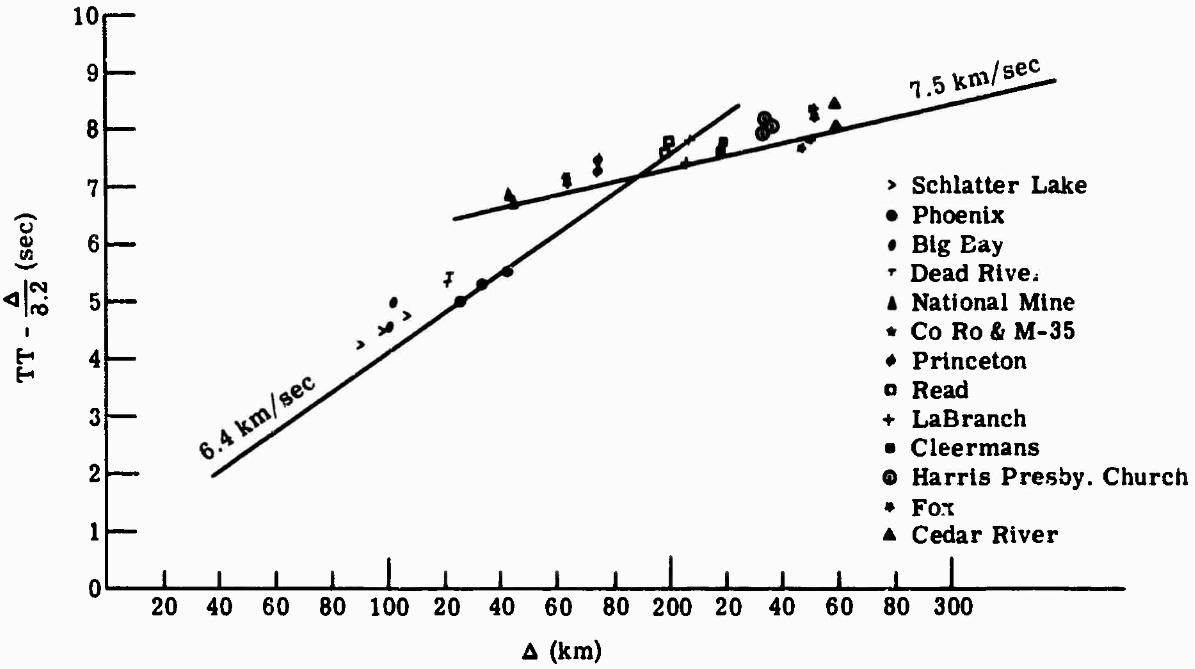

FIGURE 10. TRAVEL-TIME RESIDUALS FOR 1963 LAKE SUPERIOR SOUTH LINE.

of the October 1964 series were utilized to augment and extend the 1963 South Line to Ann Arbor, Michigan (see Willis and Tanis [1965]). These shots consisted of 1, 2, 3, 6, and 10 tons and were fired at several shot point locations. At least one shot of each size was detonated near the location of shot point 45 of the 1963 series. The reduced travel time graph for this series is shown in Figure 11. A crustal velocity of 6.17 km/sec and a P_n velocity of 8.29 km/sec gave a reasonable fit to these data.

When plans were made for participating in Project Early Rise it was decided to duplicate or reoccupy the stations used on the 1963 Lake Superior South Line and the 1964 October profile as well as extending the line southeastward to the Atlantic. Several reasons prompted this decision. Better control was needed to evaluate some early arrivals detected near the intersection of the crustal and mantle branches of the travel time curve; the variation in charge size of the 1964 shots made difficult the interpretation of the attenuation data; and finer station spacing was desirable across the Michigan Basin.

All of the Early Rise shots were fired in the vicinity of shot point 51 of the 1963 series. The reduced travel time data for the vertical component of the first compressional wave arrivals for the Early Rise shots recorded along the main northwest-southeast line are shown in Figure 12. An upper mantle velocity of 8.2 km/sec was used for these computations. A crustal velocity of 6.2 km veri. s the results

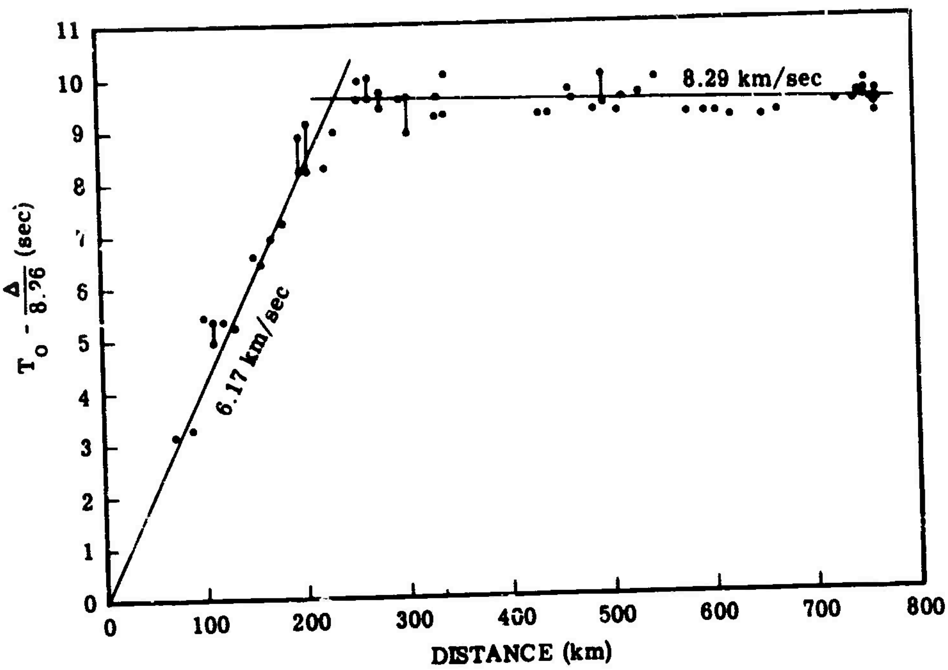

FIGURE 11. TRAVEL-TIME RESIDUALS FOR OCTOBER 1964 LAKE SUPERIOR EXPERIMENT.

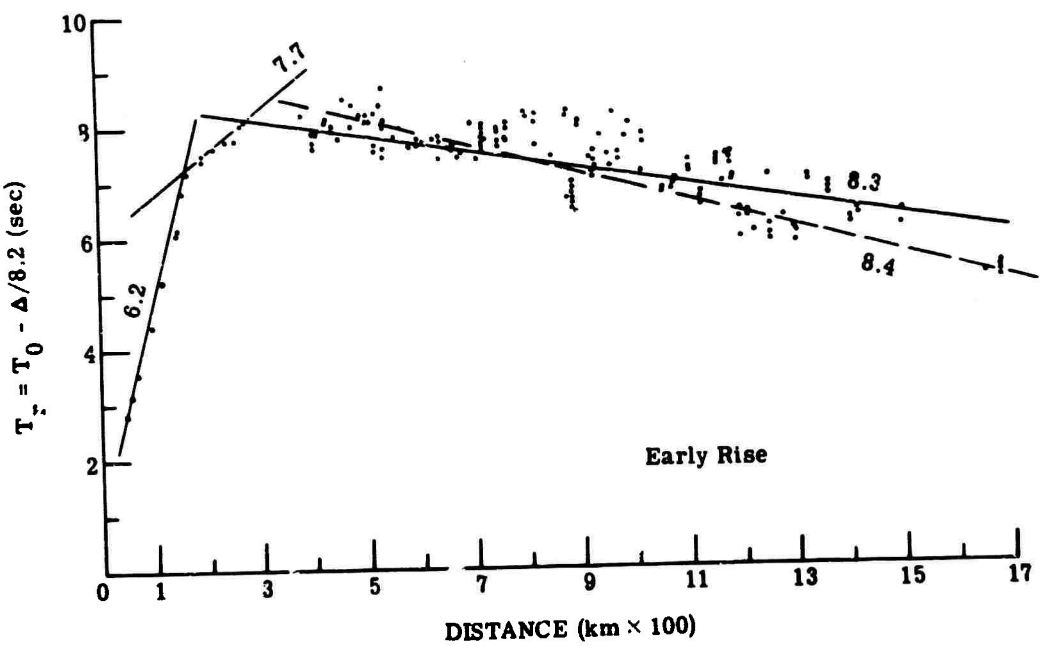

FIGURE 12. TRAVEL-TIME RESIDUALS FOR EARLY RISE.

shown in Figure 11. A clear-cut lower crustal layer with a velocity of 7.7 km/sec was obtained verifying the intermediate layer that was suspected from the data shown in Figures 10 and 11.

Considerable scatter can be seen in the P_n arrivals beyond 700 km. This is due primarily to the high seismic background noise caused by increased cultural activity across southeastern Michigan, Ohio, and West Virginia. The difficulty in determining the first arrivals along this profile can be appreciated by examining Figure 16c, page 97, of a report by Warren, et al. [1967] which shows a composite record section of this line. The details at the scale used are difficult to see. However, the increased noise beyond 750 km is quite evident. One feature particularly interesting is the small amplitude first arrivals at stations located at 160 to 275 km. This is illustrated better in Figure 13. At these distances the record gains in the field were deliberately turned up on the high gain channels so that the record would be overrecorded. This provided better record gains for the low amplitude first arrivals. This figure shows the first arrivals on the vertical component with the overrecorded later phases omitted. The records are adjusted so that arrivals with a velocity of 7.7 km/sec should be aligned vertically on this section. The top three traces are on the 6.2 km/sec branch of the travel time curve. If the record gains used in the field were set so the later phases would not be overrecorded, this early arrival would

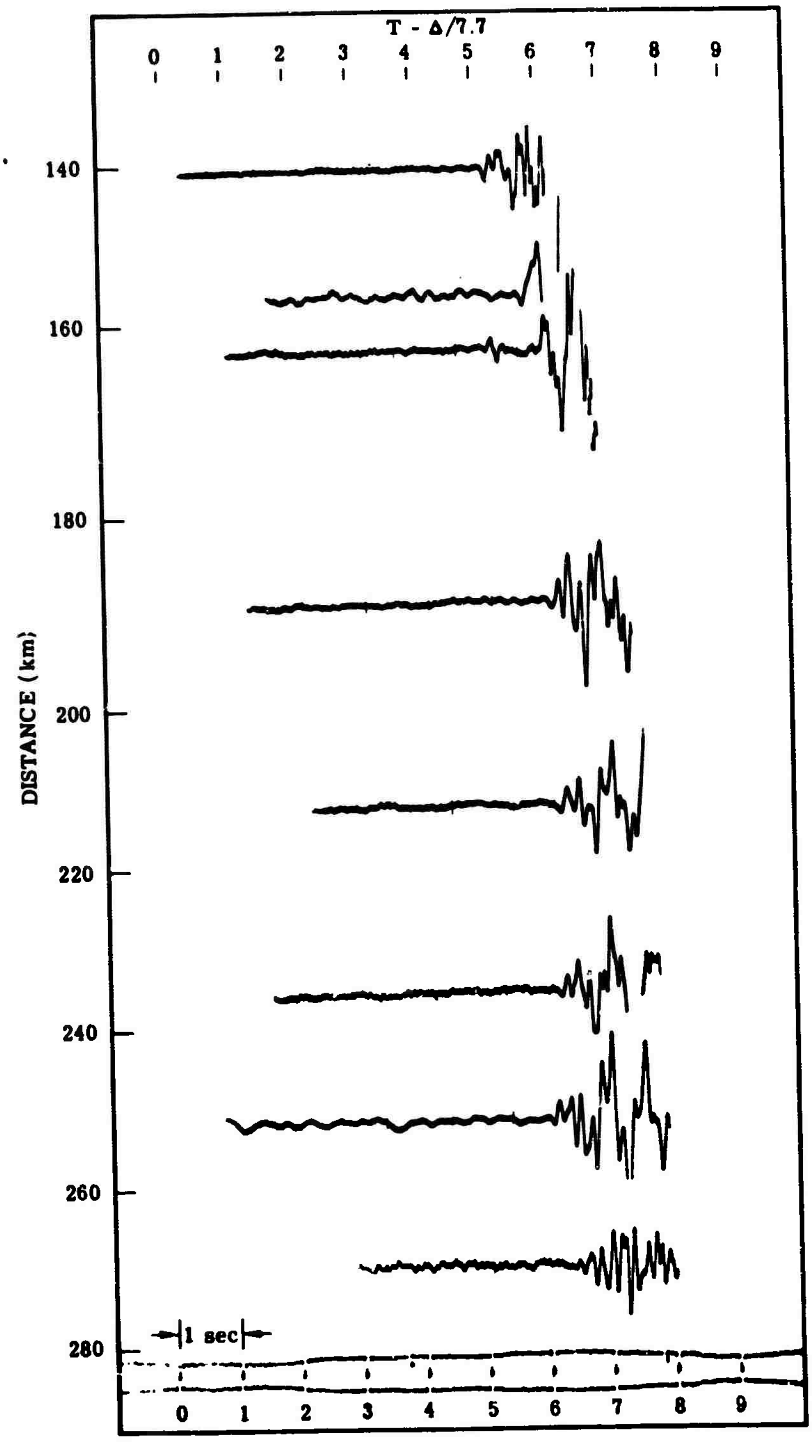

FIGURE 13. COMPOSITE SEISMOGRAM SHOWING 7.7 km/sec ARRIVAL

be very difficult if not impossible to detect. There is a second recording of another shot at each of the stations shown here. These second shots make almost perfect overlays with those shown in this figure so there can be no doubt about the validity of the apparent intermediate layer. This layer correlates with the Conrad discontinuity reported in Europe.

Figure 14 is a composite tracing of the vertical component first arrivals for all of the Early Rise shots recorded at the University of Michigan Botanical Gardens intermediate-depth well station. The record gain used (in db) for each shot is shown along the right hand margin of the figure. The repeatability of the wave form is quite good. Varying background noise prior to the first arrivals would make the identification of the first arrival difficult in some cases where only one or two recordings were available. For this reason at some of the more distant locations four or more shots were recorded in an effort to increase the accuracy of the arrival times and ground amplitudes.

The evidence for an Appalachian Mountain root system disclosed on the ECOOE Southern Profile is not as convincing on this profile. The mountains are at a distance of 1175 to 1360 km from the shot point. The residuals for those stations located on the western half of the mountains are smaller than those on the center and eastern portion and for those stations located just to the east of the mountains. However, the variation in the data closer to the shot point would

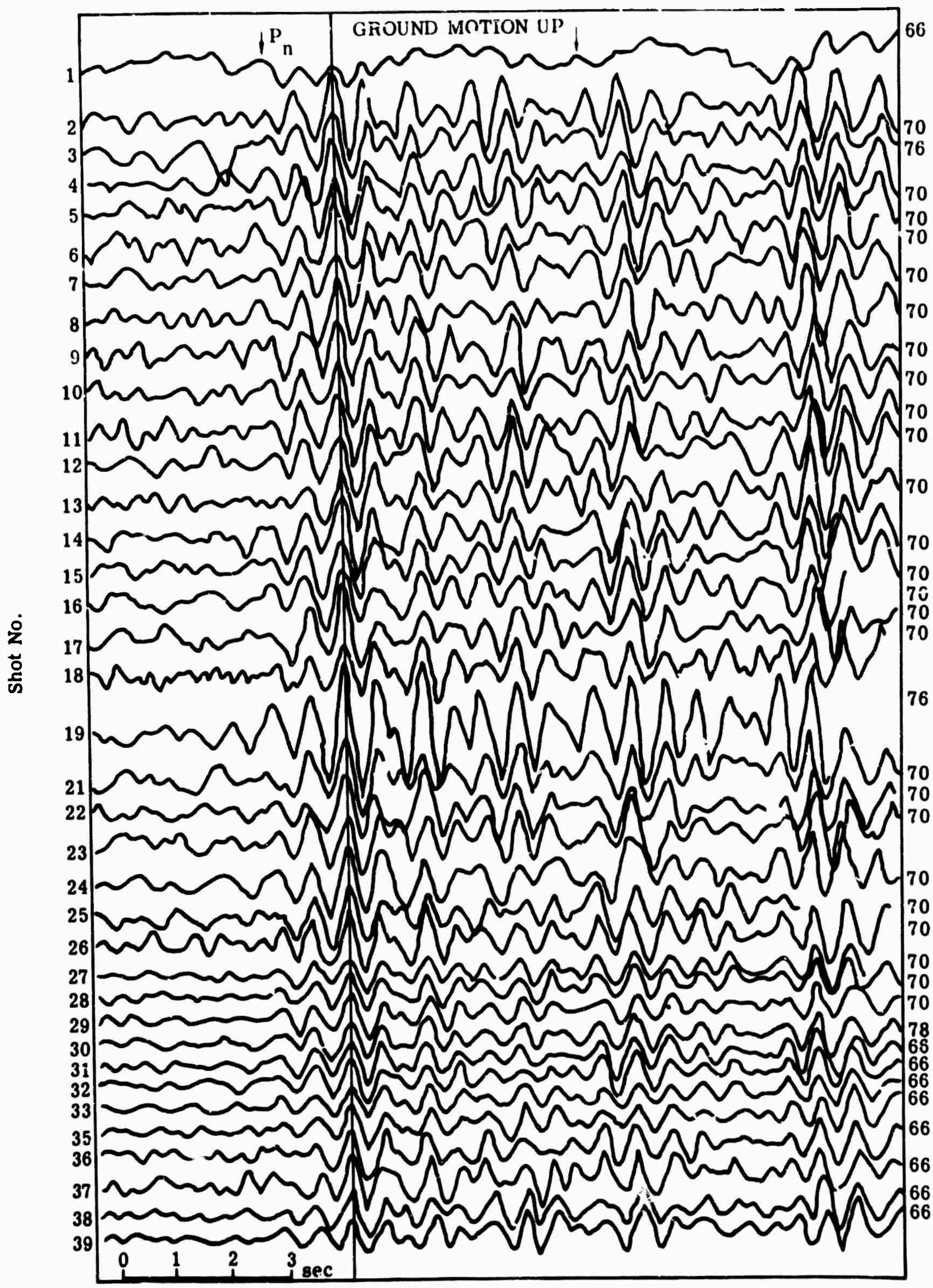

FIGURE 14. COMPOSITE SEISMOGRAM OF EARLY RISE SHOTS RECORDED AT ANN ARBOR.

cause one to think this anomaly could also be due to just scatter in the data. For the shots recorded at distances of 1500 to 1650 km, the first arrival could not be read accurately. At our most distant station the signal amplitude was larger even though the background noise was high. The vertical component recordings for four shots recorded at this furthest station are shown in Figure 15. These four individual shots were dubbed onto one magnetic tape taking into account shot zero times and slight differences in shot point locations. The times were aligned so that each shot was within ±2 milliseconds at the time of the first arrival. The maximum deviation over the entire record is ±5 milliseconds. The vertical components shown here were filtered on playback with a 1 to 4 cps passband which greatly improved the signal-to-noise ratio. The summed vertical, longitudinal and transverse components for the four shots are also shown in this figure. At this distance it can be seen that most of the energy of the first compressional wave is recorded on the vertical component. The summing technique (type of array processing) and bandpass filtering gives a 12 db improvement in signal to noise for the first arrival. Such techniques allow a more accurate determination of the first arrival times but it is quite laborious to make the composite magnetic tape dubs. This process was used only where control points were needed for critical interpretations and where four or more shots were recorded at the same station.

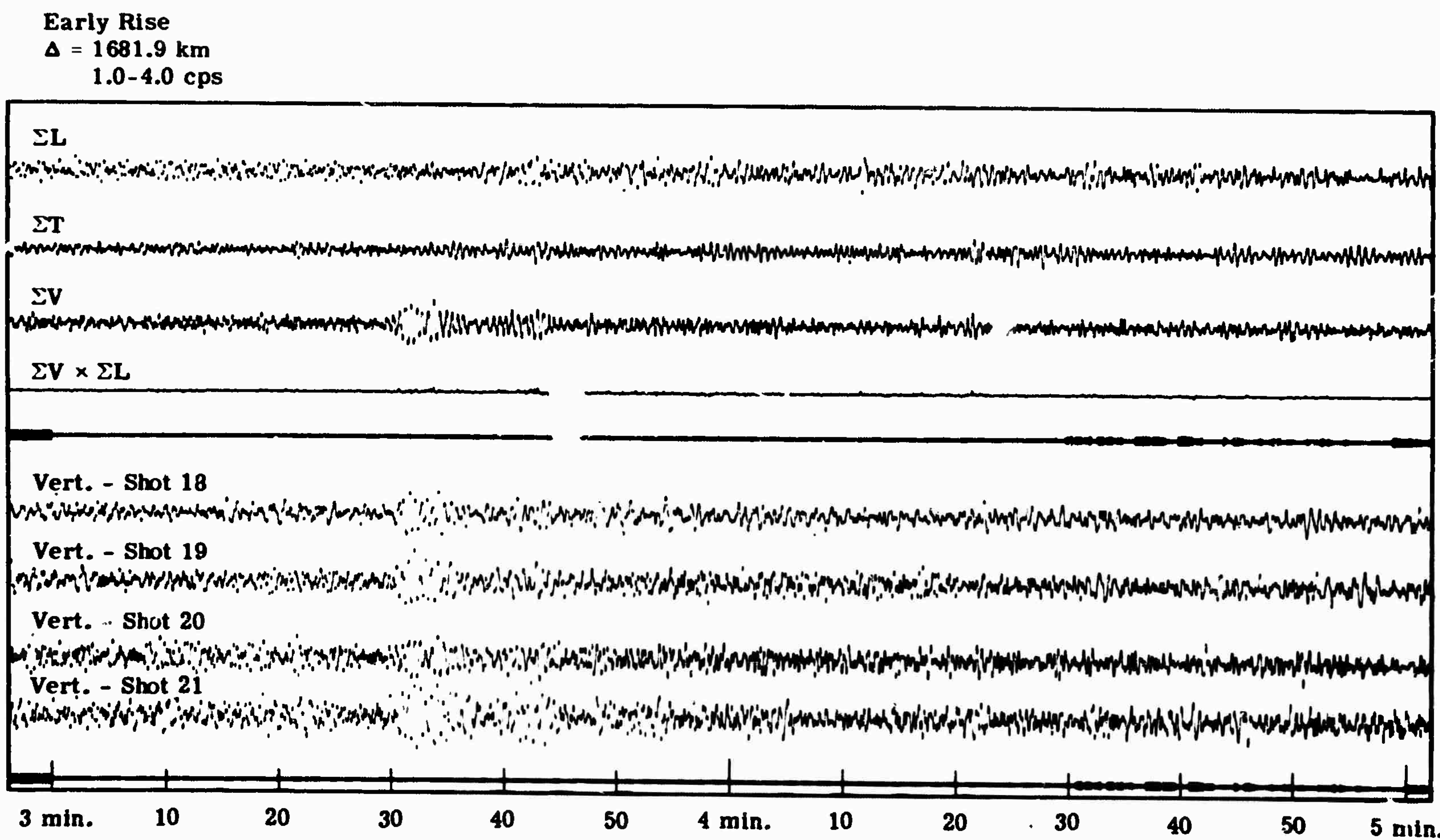

FIGURE 15. EARLY RISE SHOTS 18-21 RECORDED AT POTTER'S HILL, NORTH CAROLINA

The combined travel time data of Figures 10-12 yield the following:

$T_1 = \Delta/4.8^* + 0$	$h_1 = 2.9$	(18)
$T_2 = \Delta/6.2 + 1.0$	$h_2 = 22.5$	(19)
$T_3 = \Delta/7.7 + 5.9$	$h_3 = 25.0$	(20)
$T_4 = \Delta/8.3 + 8.5$	$h_4 = 21.6$	(21)
$T_5 = \Delta/8.4 + 9.3$		(22)

*From Smith, et al. (1966)

A comparison of the crustal and upper mantle structure at both ends of this long range refraction profile shows considerable differences. The total crustal thickness at the continental margin is 28.1 km compared to 50.4 km at Lake Superior. Two major crustal layers with velocities of 6.2 and 7.7 km/sec are clearly evident at the latter area while only a single layer with a velocity 6.0 km/sec was found at the continental margin. A thin layer of low velocity sediments at the surface is assumed in both areas. A slightly higher upper mantle velocity (8.3 vs 8.2 km/sec) was found in the Lake Superior area while the higher velocity layer in the mantle was found at a shallower depth (72 km vs 99 km at the continental margin). However, the poor signal-to-noise ratios and the scatter of the data make the validity of the latter somewhat questionable.

If the data shown in Figure 11 were replotted in Figure 12 using the same upper mantle velocity of 8.2 km/sec for computing the residuals, the P_n branch of the travel time residuals for the October 1964 shots would average about 1

second later than those of the Early Rise shots. The 6.2 km/sec branch agrees quite well while the 7.7 km/sec branch is approximately 0.3 seconds late. This could only be explained by a difference in crustal structure between shot points 45 and 51. Figure 22 in Smith, et al. (1966) shows that shot point 45 is located on the flank of an anomalous rise in the surface of the upper mantle and that a trough or depression in the mantle is apparent between these two shot points. The observed discrepancy of one second would indicate a difference in crustal thickness of the order of 6 or 7 km. The mantle depression reported by Smith, et al. (1966) is not quite that large. However, a ray traversing this feature obliquely would not require the depression to be as deep as 6 km to cause the one second travel time discrepancy.

Many other authors (Smith, et al. [1966], Berry and West [1966], Mereu [1966], Cohen and Meyer [1966], and others) report a single layer crust with a high velocity (6.5 to 7.2 km/sec). Our Copper Harbor East and West profiles for the 1963 Lake Superior Experiment showed only one crustal layer with a velocity of 6.60 to 6.71 km/sec. However, the Knife River, Minnesota Profile disclosed two crustal layers with velocities of 6.19 and 7.01 km/sec. Green and Hales [1967] and Roller and Jackson [1966] report a two-layer crust on their southwestern profiles from Lake Superior with velocities of 6.3/5.5 and 6.97/7.2 km/sec respectively. Why then should our profile extending southeastward from the Keweenaw

Peninsula show the two-layer crust while our other two Keweenaw Peninsula profiles indicate only one? The simplest explanation lies with the fact that the onset of the 7.7 km/sec arrival has a very small amplitude (a factor of 30 lower than the later arrival corresponding to 6.2 layer). This lower amplitude event could go undetected because where only one record gain is used, that gain would be set so as not to overrecord the later arrival. We used double or triple gains to investigate this anomaly. For the Early Rise shots we used close station separations at distances of 140 to 280 km to verify the suspected two-layered crust. The fact that the same shot point was used each time also greatly facilitated this study since the small amplitude first arrivals could be identified by overlaying the records recorded at the same stations. Such was not possible for the 1963 series.

The total thickness of the crust is not altered significantly (only 0.1%) by the two-layered crust with velocities of 6.2 and 7.7 km/sec compared to a single-layered crust of 6.8.

Attenuation Studies

The detonation of large high explosive shots in several hundred feet of water provides a very efficient coupling mechanism for seismic waves (see Weston [1960], DeNoyer and Willis [1961] and Willis [1963]). The increased signal level for a properly tamped water shot can generate ground displacements that are a factor of 20 or more larger than

an equivalent sized shot fired in rock. Hence, the firing of 1 to 10 ton shots in Lake Superior and off the Atlantic Coast has provided seismologists with a source that can be detected to ranges on the order of 2500 km under favorable conditions.

Earlier research (Willis, 1964) had shown that the attenuation of high frequency (above 5 cps) seismic energy at distances of several hundred kilometers was significantly lower in the northeastern United States than in the western areas of the country. Hence recording equipment was used that would insure recording any high frequency energy that might be generated by these underwater shots.

ECOOE Series

The observed loss in amplitude as a function of distance from the 1962 North Carolina shots, the ECOOE series, the CHASE shots, and the 1963-64 Lake Superior shots has been reported in nine previous papers by the author. In Figures 15-18, 21, 24, and 27 of Willis and Jackson [1966] the vertical component of the first compressional wave arrival for the ECOOE shots can be seen to fall off like the inverse cube of the distance. Figures 22 and 25 likewise show a similar fall off for the maximum amplitude P wave arrival. The maximum surface-wave arrival shown in Figures 23 and 26 can be seen to fall off at a rate between the inverse square and the inverse cube of the distance.

P_n attenuation curves for the individual 5-ton shots of the ECOOE northern profile are shown in Figure 16. The major fluctuation in amplitude correlates with those stations in the Appalachian Mountains. The signal levels of the vertical component of first compressional wave (P) arrival and the maximum surface wave (R) arrival are shown in Figure 17 for the ECOOE 10-ton shots (labeled 6 and 39) of the southern profile and for CHASE III and IV (labeled) 3 and 4). The rate of attenuation is approximately proportional to the inverse cube of the distance. An anomalous increase in signal with distance was observed across the Michigan Basin. It is also interesting to note that the maximum surface wave amplitudes for the 10-ton shots (6R and 39R) are of the same order of magnitude at many distances as the compressional wave amplitudes for CHASE III and IV. For the same shot the maximum amplitude surface waves ranged 10 to 20 db higher than the first arrival compressional waves.

October 1964 Lake Superior Shots

Attenuation studies of this series of recordings are complicated by the fact that the charge size varied from one to ten tons and multiple shot points were used. Comparing recordings made for the same shot, disclose anomalous amplitude variations such as that shown in Figure 18. Shown here are the vertical component ground particle velocity amplitudes of the first compressional wave arrival for five stations ranging in distance from 549 km to 766 km. The two most distant stations had larger peak amplitudes than the

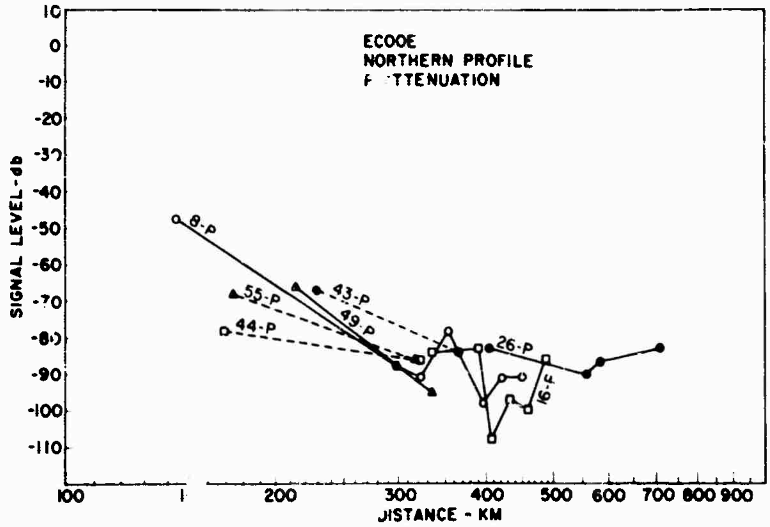

FIGURE 16. SIGNAL-LEVEL vs DISTANCE FOR 5-TON SHOTS OF ECOOE NORTHERN PROFILE.

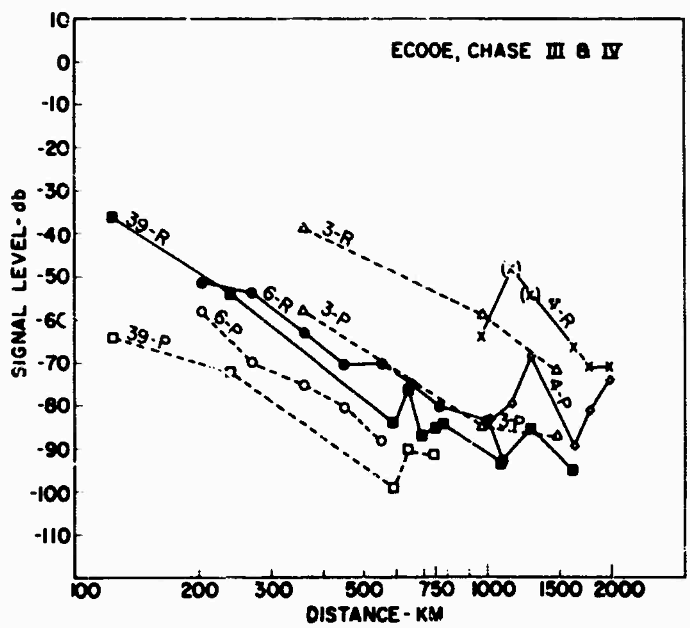

FIGURE 17. SIGNAL-LEVEL vs DISTANCE FOR 10-TON SHOTS OF ECOOE SOUTHERN PROFILE AND CHASE III AND IV.

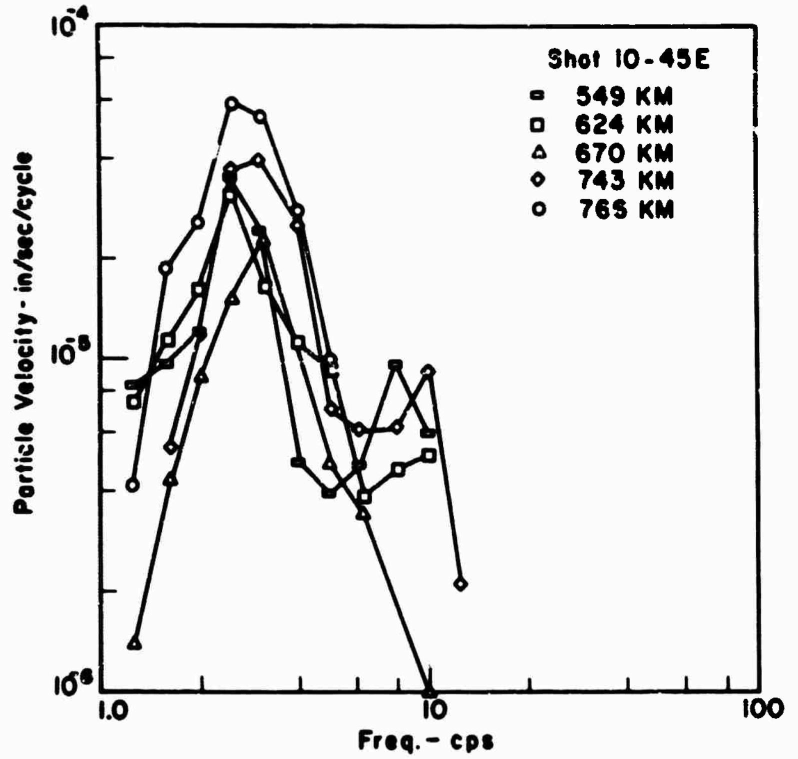

FIGURE 18. SPECTRAL ANALYSIS OF 10-TON LAKE SUPERIOR SHOT RECORDED AT STATIONS LOCATED IN MICHIGAN BASIN.

close stations and the center station had the lowest amplitude. Some of these variations are probably due to local geological factors. However, all of the stations are located in the same type of geological environment (southeast flank of the Michigan Basin) and are spaced relatively close together.

In order to determine a mathematical expression that would best fit all of the recorded first compressional wave data from this experiment a digital computer program was written that would compute the essential variables. In order to eliminate shot point effects, only those recordings from shots fired at shot point 45 were used.

We assume* that

$$A_i = A_o \left(\frac{x_i}{x_o}\right)^{-n} e^{-\alpha f(x_i - x_o)} \tag{23}$$

$$\frac{A_i}{A_j} = \left(\frac{W_i}{W_j}\right)^{N}$$

where A = particle velocity amplitude

x = distance

n = geometrical spreading factor

α = absorption coefficient

f = frequency

W = charge size

Solving for α, we have

---

*See Willis [1960], Willis and Wilson [1960].

$$\alpha_{ij} = \frac{\ln\frac{A_i}{A_j} + n\ln\frac{x_j}{x_i} + N\ \ln\frac{W_j}{W_i}}{f(x_j - x_i)} \qquad (24)$$

In the computer program N was allowed to range from 0.2 to 2.0 in increments of 0.1. Values of α less than 10^{-4} or greater than 10^{-1} were omitted. The set of alphas $\{\alpha_{ij}: i<j\}$ were computed for each value of N. The alphas were rounded off to the first significant figure. Some of the results of these computations are shown in Figures 19 through 22. In Figure 19 n was set equal to 1 and the frequency used was 3.15 cps which corresponds to the predominant peak observed at most distances. Each point $P(\alpha,N)$ represents the number of times a particular value of α occurred for a particular value of N. The actual numbers have been omitted here for the sake of clarity but the contours show the range in values. Using the same geometrical spreading factor similar graphs were made for f = 1.6, 2.0, 2.5, 4.0, 5.0, 6.3, 8, and 10 cps. The narrow ridge of values centering at $\alpha = 2 \times 10^{-3}$ was found to shift slightly with frequency. At 8 to 10 cps, larger values were observed for N>1 and $\alpha = 1 \times 10^{-3}$ while for f = 1.6 to 2.0 cps larger values of P were found for N<1 and $\alpha = 3 \times 10^{-3}$. The elongated ridge shown in Figure 19 was somewhat surprising since it was thought that P would have a pronounced peak in the range of N between 0.7 and 1.0 (Willis and Turpening, 1963).

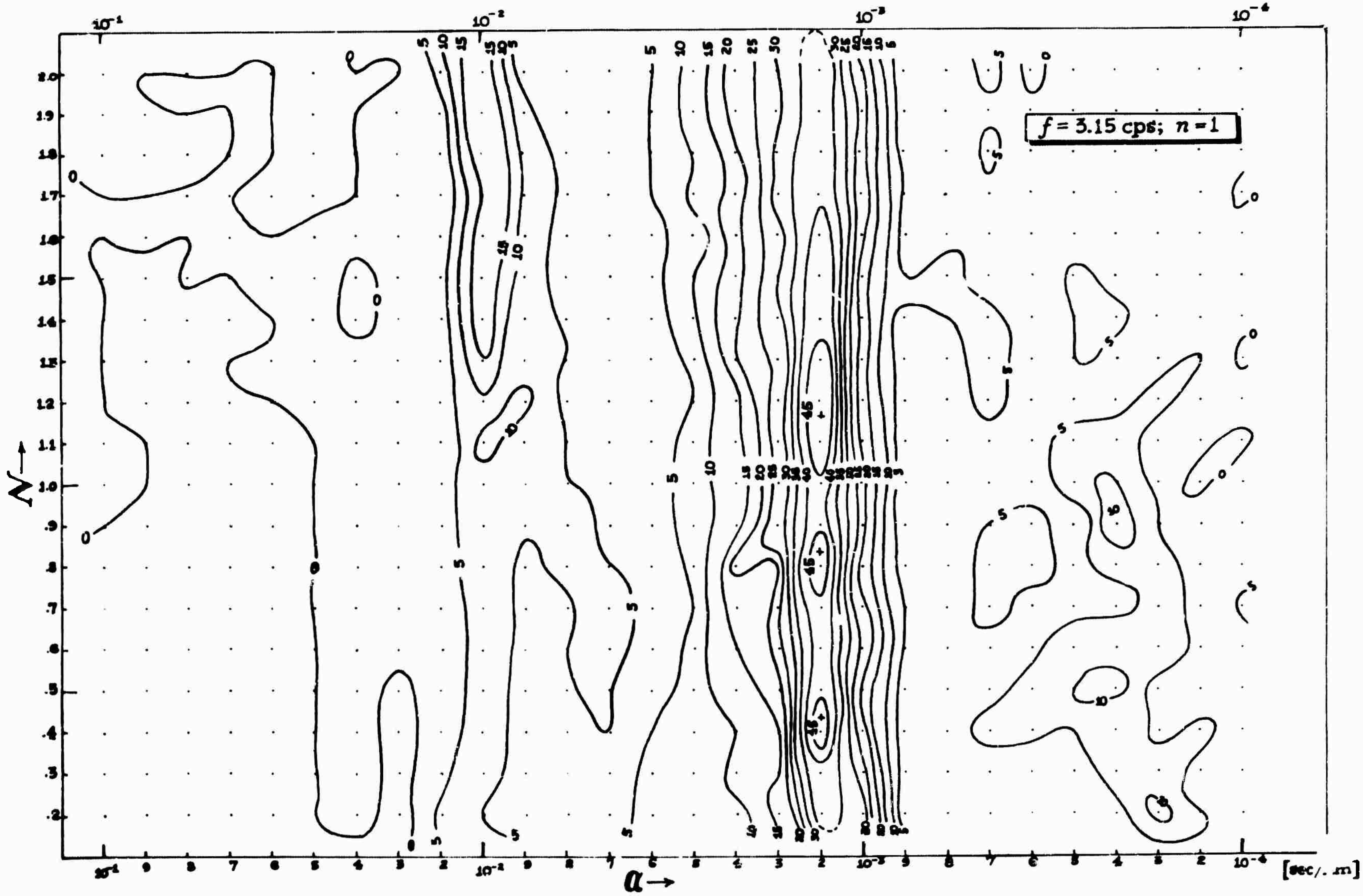

FIGURE 19. 1964 LAKE SUPERIOR ATTENUATION DATA: α vs N, f = 3.15 cps AND n = 1.

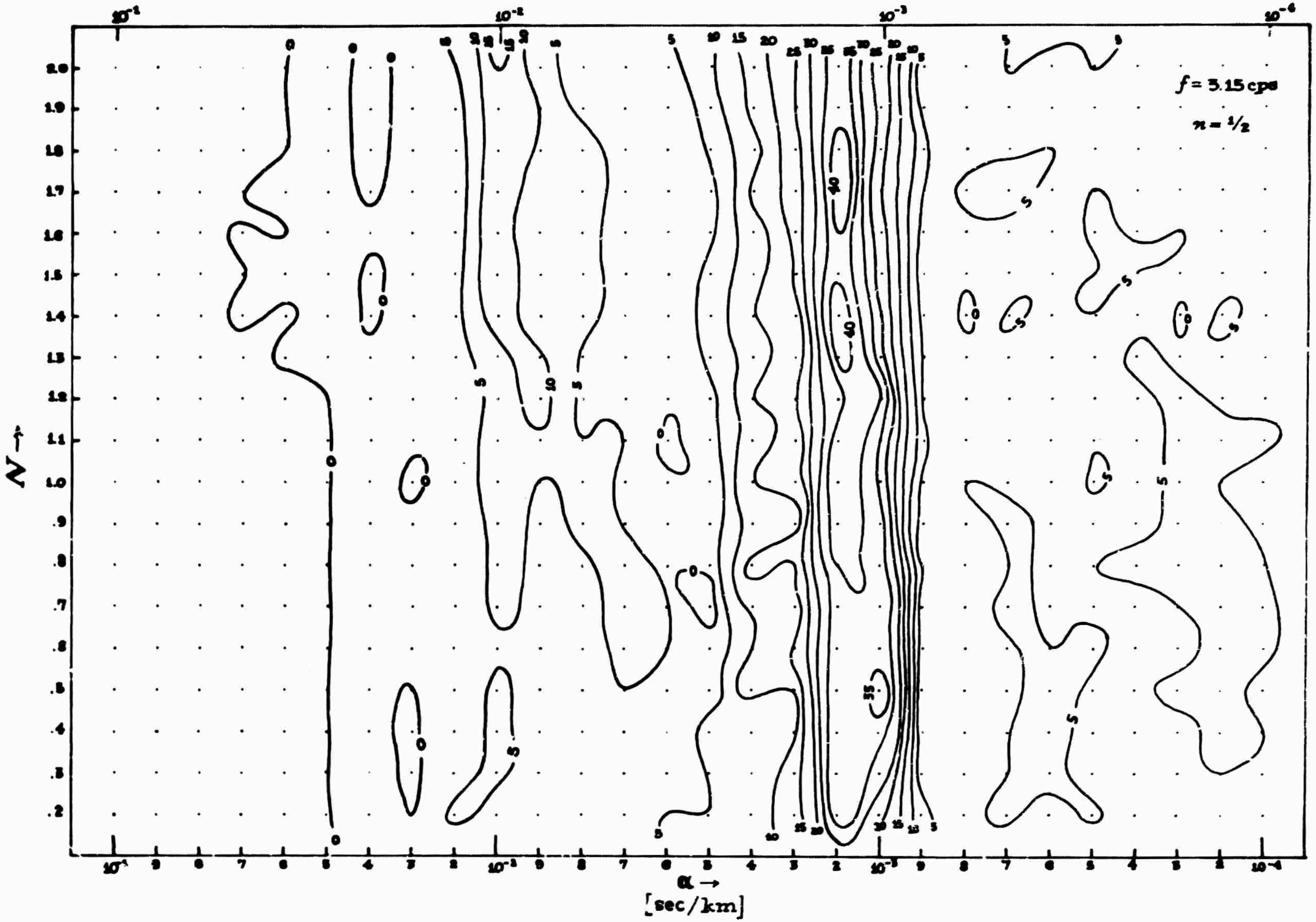

FIGURE 20. 1964 LAKE SUPERIOR ATTENUATION DATA: α vs M. f = 3.15 cps AND n = 0.5.

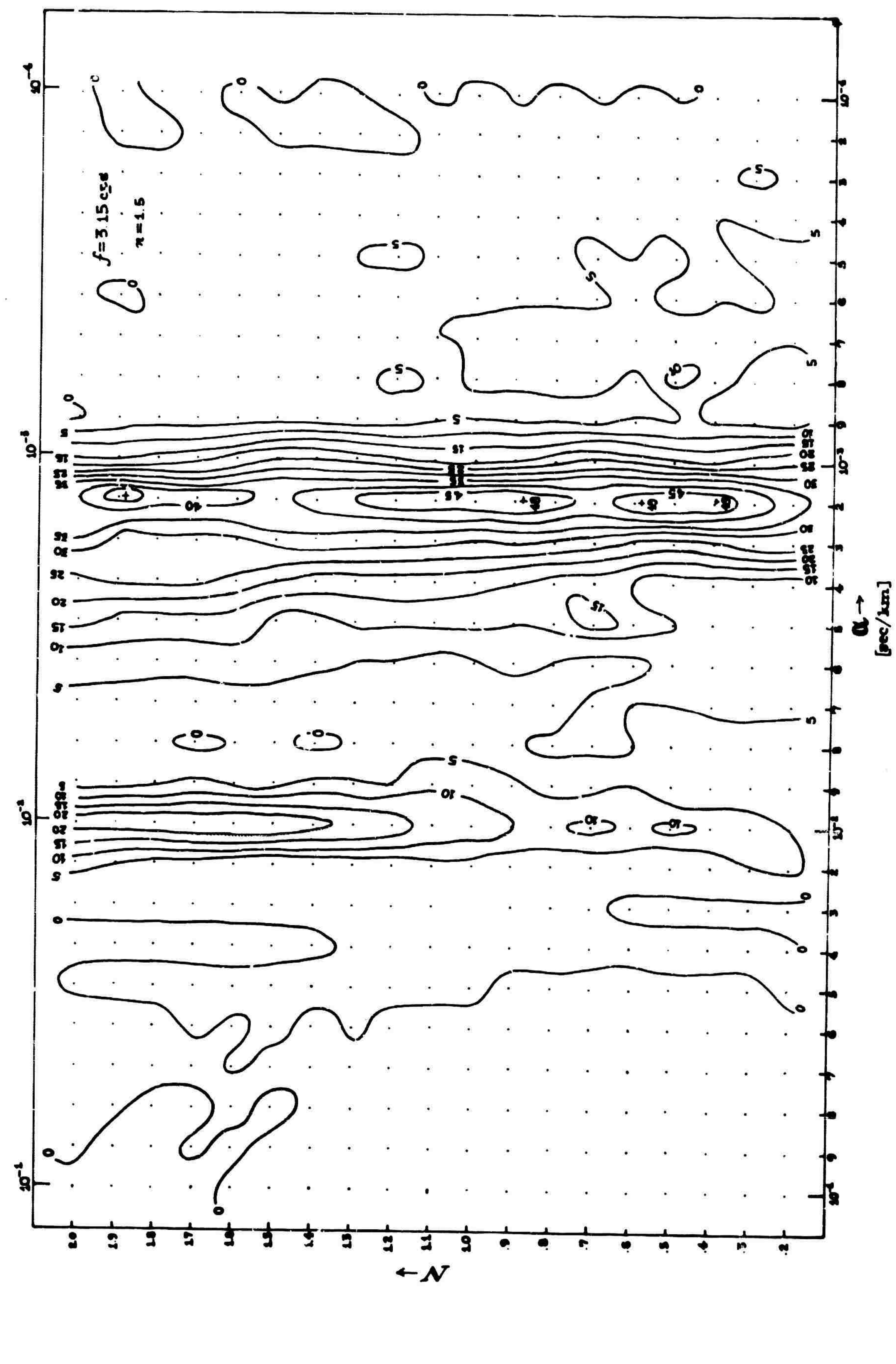

FIGURE 21. 1964 LAKE SUPERIOR ATTENUATION DATA: α vs N, f = 3.15 cps and n = 1.5.

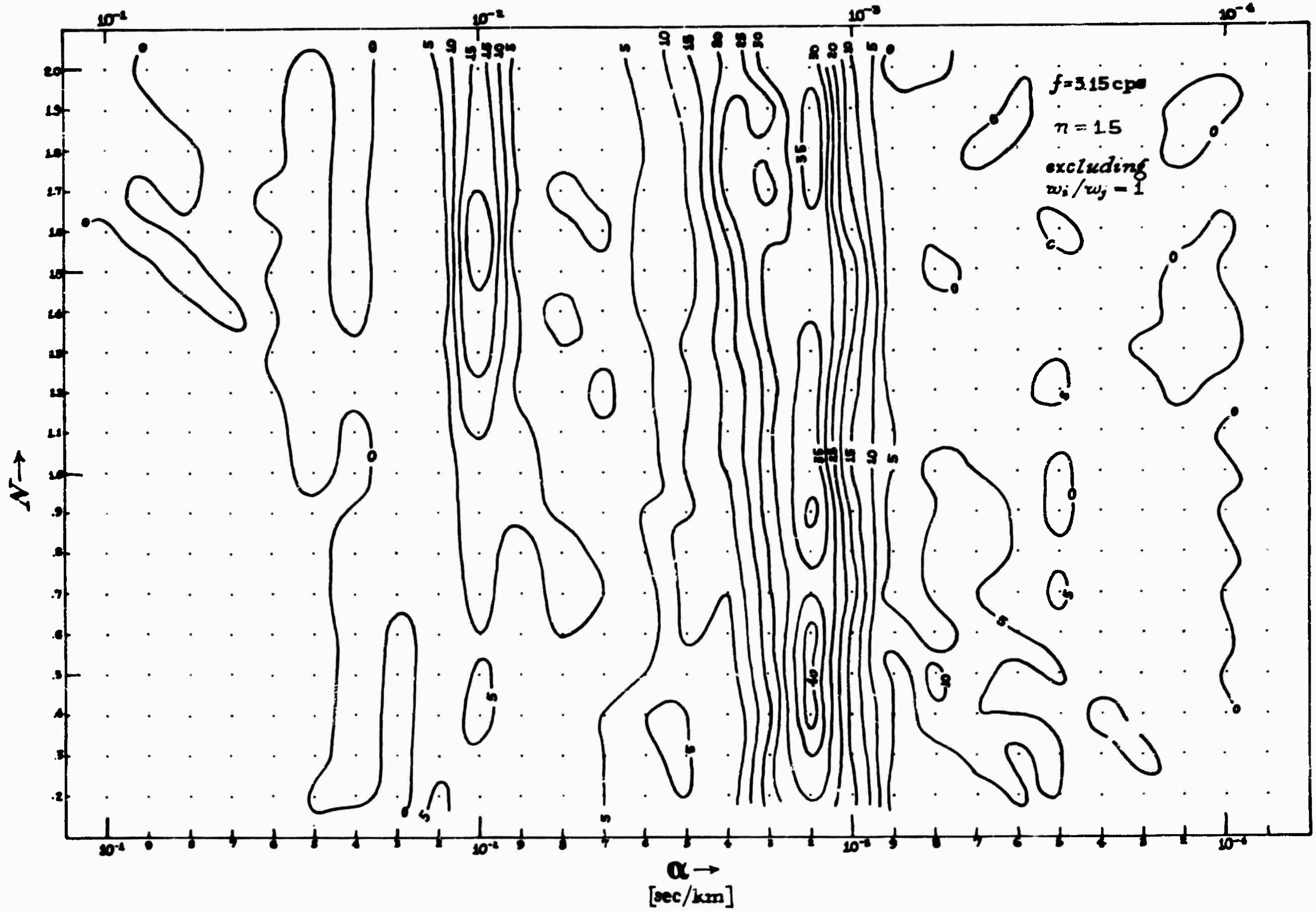

FIGURE 22. 1964 LAKE SUPERIOR ATTENUATION DATA: α vs N, f = 3.15 cps, n = 1.5, EXCLUDING $\omega_i/\omega_j = 1$.

Similar graphs were prepared for $n = 1/2$, $3/2$, and 2 at $f = 3.15$ cps. Two of these are shown in Figures 20 and 21. For the four values of geometrical spreading the largest values of P occurred at $n = 3/2$. However, for $n = 1$, P was not significantly smaller. A secondary peak in α occurs at $\alpha = 10^{-2}$. One would intuitively associate this with the crustal attenuation. However, these larger values of α are found primarily in association with those sets of stations where the station separation is small ($\Delta \simeq 25$ km). In those cases where $W_i = W_j$, N would have no significance. In order to determine if this were the reason for the elongated ridge at $\alpha = 2 \times 10^{-3}$, all sets of stations where $W_i/W_j = 1$ were omitted. The results are shown in Figure 22 for $n = 3/2$ and $f = 3.15$ cps. It can be seen that there is no significant change in pattern from that shown in Figure 21. A weighting factor based on the percentage of path length through the upper mantle compared to the total path length was used for the data shown in Figure 21. The significant features remained unchanged. Hence, varying the charge size added an additional parameter that seriously interfered with the interpretation of the attenuation of the seismic waves generated by this series of shots. This could have been minimized if one station had been operated at a small epicentral distance to record all of the shots at the same point. In this manner the shots could have been normalized to a standard reference and corrections made accordingly.

Early Rise Shots

Figure 23 shows the first arrival vertical component of ground displacement versus distance for the Early Rise data. While there is considerable scatter, the data show an average attenuation proportional to the inverse 3/2 power of the distance. On the reversed profile, the ECOOE shots, the 1962 North Carolina data and CHASE shots fit more closely to an inverse third power relationship of the distance. Furthermore, for equivalent sized shots, the shots fired in Lake Superior were more efficient in coupling energy into seismic waves than the shots fired in the Atlantic Ocean. The signal amplitudes ranged from 6 to 18 db or a factor of 2 to 8 larger for shots recorded at comparable distances. A partial explanation for this phenomenon may be that the shots fired in Lake Superior (especially the Early Rise data) were fired at an optimum depth for signal enhancement due to reinforcement by the bubble pulse (optimum depth between 593 and 607 feet).

Attenuation measurements made in the western United States (Herrin and Taggart [1962], Jordan, et al. [1965], Nuttli [1963], Romney [1959], and others) indicate a fall off in amplitude proportional to the inverse cube of the distance. Considerable asymmetry was also disclosed, especially in connection with the GNOME shot. Profiles west and northwest from Carlsbad, New Mexico showed small amplitude first arrivals arriving several seconds late while to the northeast large first arrival amplitudes could be correlated with early arrivals. It is postulated that a low velocity layer in the

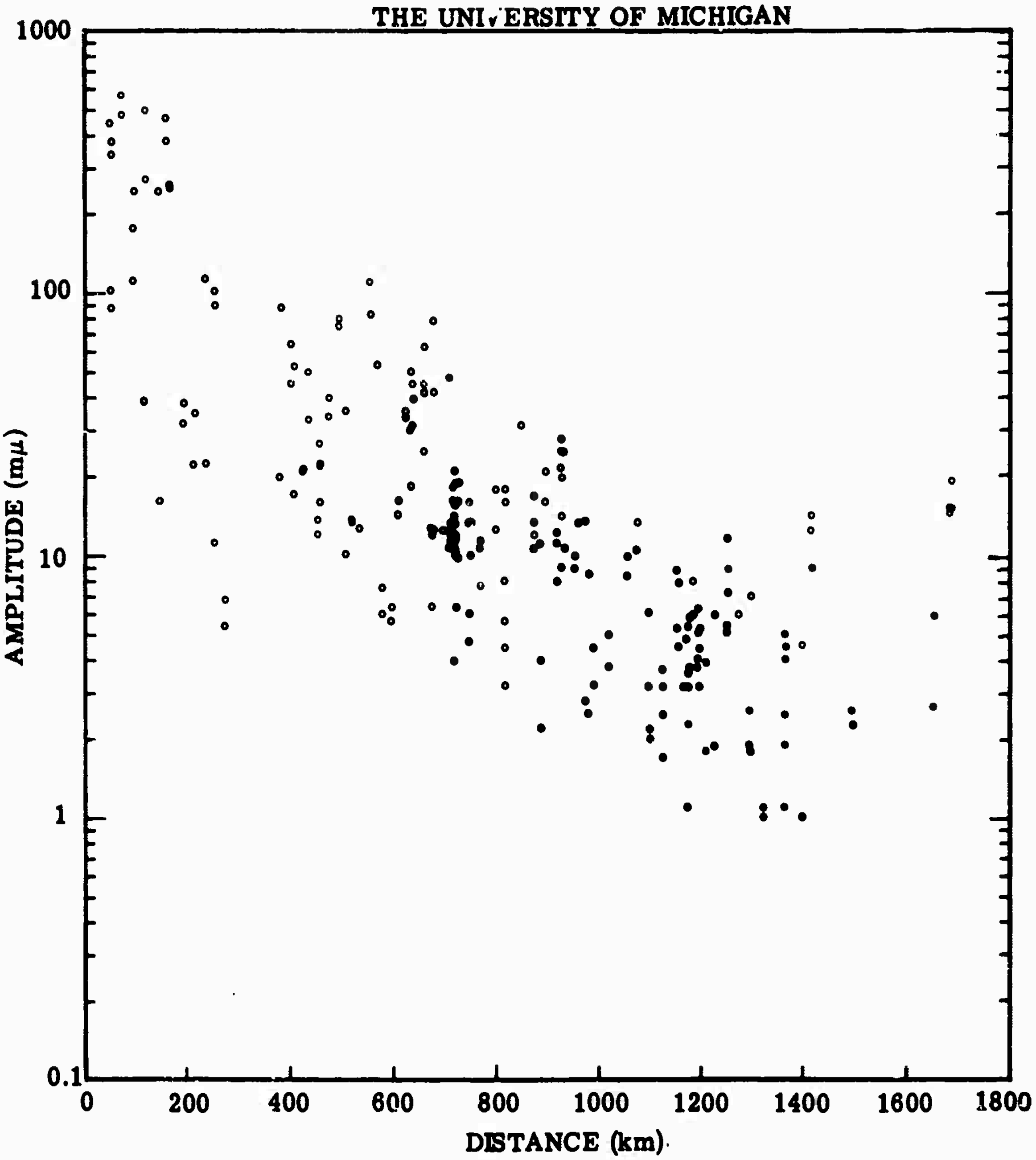

FIGURE 23. GROUND DISPLACEMENT vs DISTANCE FOR PROJECT EARLY RISE.

upper mantle in the western United States could account for this anomaly. The Lake Superior shots verified that there was no low velocity layer in the central or eastern United States. However, the shots fired in the Atlantic display the inverse cube relationship of amplitude fall off with distance.

Iso-Particle Velocity Maps

Display of seismic attenuation data in a reasonable manner has been a difficult problem in the past. Presentations such as shown in the former section are restrictive in that only the peak signals are presented and the frequency content is unknown. On the other hand detailed frequency analysis such as shown in Figure 18 are difficult to follow since the spectral curves overlap. The apparent scatter of the data in this form makes the interpretation very difficult. Changes in the spectra and rapid changes in amplitude between stations contribute to this difficulty. A method of presentation of attenuation data was developed (Willis, et al 1964, Willis and DeNoyer, 1966) that permits presentation of the attenuation data in three dimensional distance, frequency and particle velocity space. Contour maps are constructed where a linear distance scale and a logarithmic frequency scale are used. The particle velocity values are contoured at 6 db intervals (a factor of 2) using a reference of 1×10^{-2} in/sec/cycle. These amplitudes correspond to true ground motion. All of the instrumentation response has been removed.

Figure 24 shows the vertical component of the ground particle velocity for the first three cycles of the first compressional wave arrival recorded on the Early Rise shots. The gap in data from 270 to 380 km is where the profile crosses Lake Michigan. The dots correspond to data points (the values have been omitted for sake of clarity). The spectra of these shots are strongly influenced by the bubble pulse frequency and the reverberation of the sound waves between the lake bottom and the surface. The bubble pulse frequency is approximately 2 cps and would be in phase at that water depth. The organ pipe frequencies for constructive and destructive interference are:

Constructive	Destructive
f_0 = 2 cps	$2\ f_0$ = 4 cps
$3\ f_0$ = 6 cps	$4\ f_0$ = 8 cps
$5\ f_0$ = 10 cps	$6\ f_0$ = 12 cps
$7\ f_0$ = 14 cps	

A rather remarkable correlation with these frequencies can be seen between the particle velocity high's and low's shown in this figure.

The major peaks along the entire profile occur at approximately 2 and 5-6 cps. The former coincides with the bubble pulse frequency and the fundamental organ pipe mode while the latter coincides with the third harmonic. A high can also be correlated quite well with the 5th harmonic. The particle velocity lows correlate quite well with the second and fourth harmonic where destructive interference

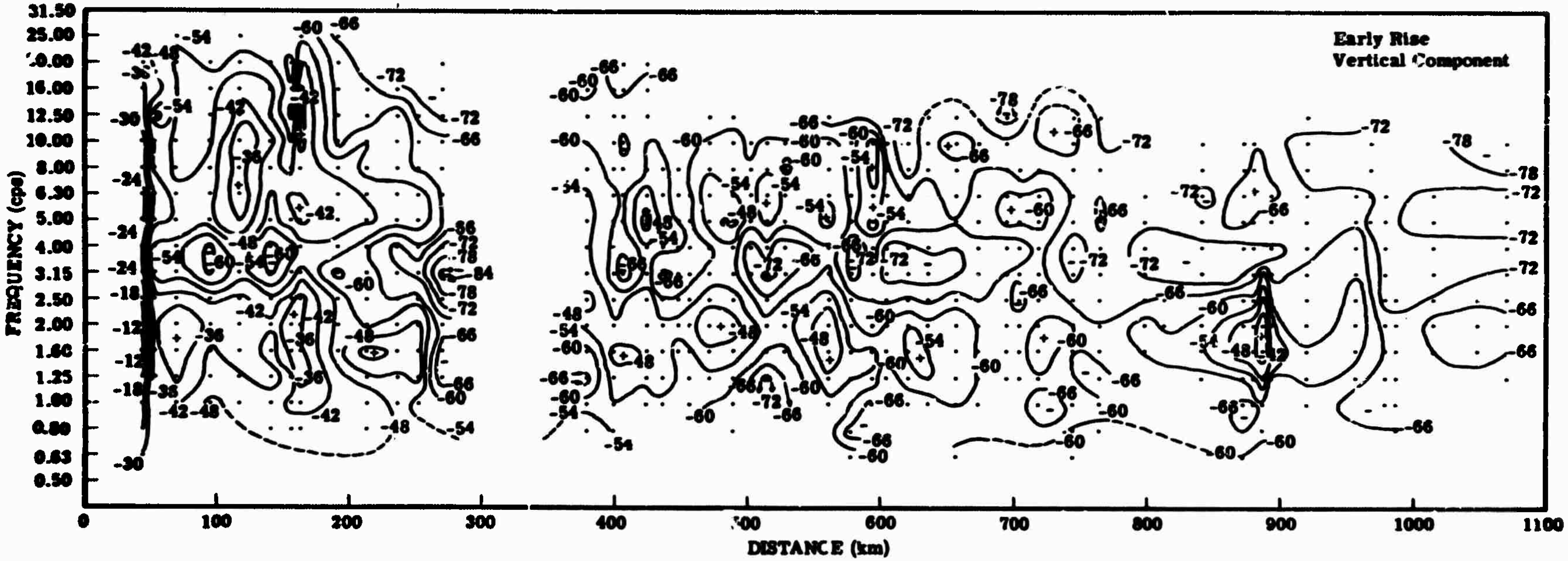

FIGURE 24. ISO-PARTICLE-VELOCITY MAP OF FIRST P-WAVE ARRIVAL FOR EARLY RISE SHOTS.

should occur.

A particle velocity low at all frequencies can be seen near the crossover point (160 km) of the 6.2 and the 7.7 km/sec branch of the travel time curves. Unfortunately, the P_n crossover point falls near that section of Lake Michigan where no data are available. One other predominant feature should be discussed. A pronounced particle velocity high occurs at approximately 885 km. This same high was observed at a group of four stations at the same distance located near Anna, Ohio. The latter area is 150 km southwest of the main profile and is an area where small earthquakes have been recorded in the past. The cause of this large amplitude is not known. It may possibly correlate with a cusp in the travel time curves. Its buildup is observable at seven stations.

Figure 25 shows the iso-particle-velocity map for the 1963 Lake Superior South Line. The particle velocity high's correlated well with the bubble pulse and organ pipe modes that would cause reinforcement. The predominant signal along the entire profile is 4 cps which correlates with the first odd harmonic of the organ pipe mode. The sharp decrease in particle velocity amplitude at 160 to 170 km correlates with the data on the Early Rise shot which occurred at the crossover point on the travel time curve from the 6.2 to 7.7 branch.

Other iso-particle-velocity maps of the 1963 Lake Superior Experiment were published in a paper by Willis and DeNoyer [1966]. These included compressional and shear wave vertical component maps made for data recorded on the

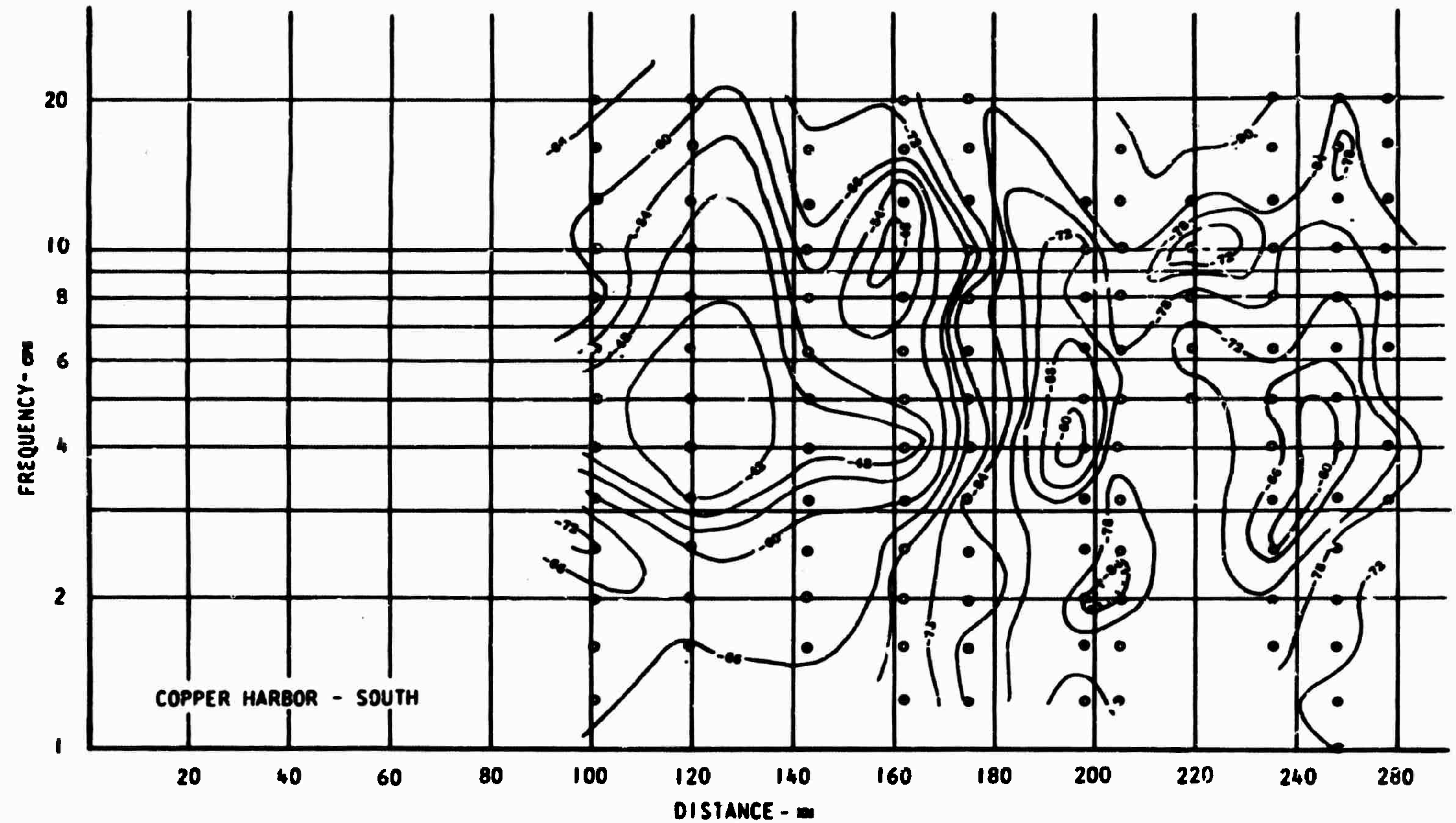

FIGURE 25. ISO-PARTICLE-VELOCITY MAP OF FIRST P-WAVE ARRIVAL FOR 1963 LAKE SUPERIOR SOUTH LINE.

Keweenaw Peninsula for shots fired in the eastern and western portions of the lake and also for the entire series of shots recorded at Knife River, Minnesota. The attenuation pattern shown in these maps is quite complex. Many of the fluctuations in spectral data persisted from shot to shot where the station remained fixed and from station to station when the shot point was fixed. These trends are difficult to follow when examining the individual spectral curves. However, there are many anomalies that can be correlated at comparable distances on each profile. Likewise anomalies can be seen on each line which correlate with the shot and are independent of the epicentral distance.

Figures 26 and 27 show the particle velocity maps for the first compressional wave arrivals for the 1962 North Carolina Experiment. The Potter's Hill and Dudley recording stations were located in line with the shot profile and were 40 km apart. The similar major features which correlate with the same shot points at these two stations prove that conditions in the source region strongly influence the character of the seismic signal recorded at distant stations. Vertical and longitudinal component iso-particle-velocity maps for the first P and the maximum P arrival were made for the ECOOE Southern Profile 150, 250, 350, 400, 450, and 550/600 km stations. Similar ma r the 150, 300, and 325 km sites on the Northern Profile were also constructed. Figure 28 shows the vertical component particle velocity amplitudes for the 150 km site on the ECOOE Southern Profile. st of

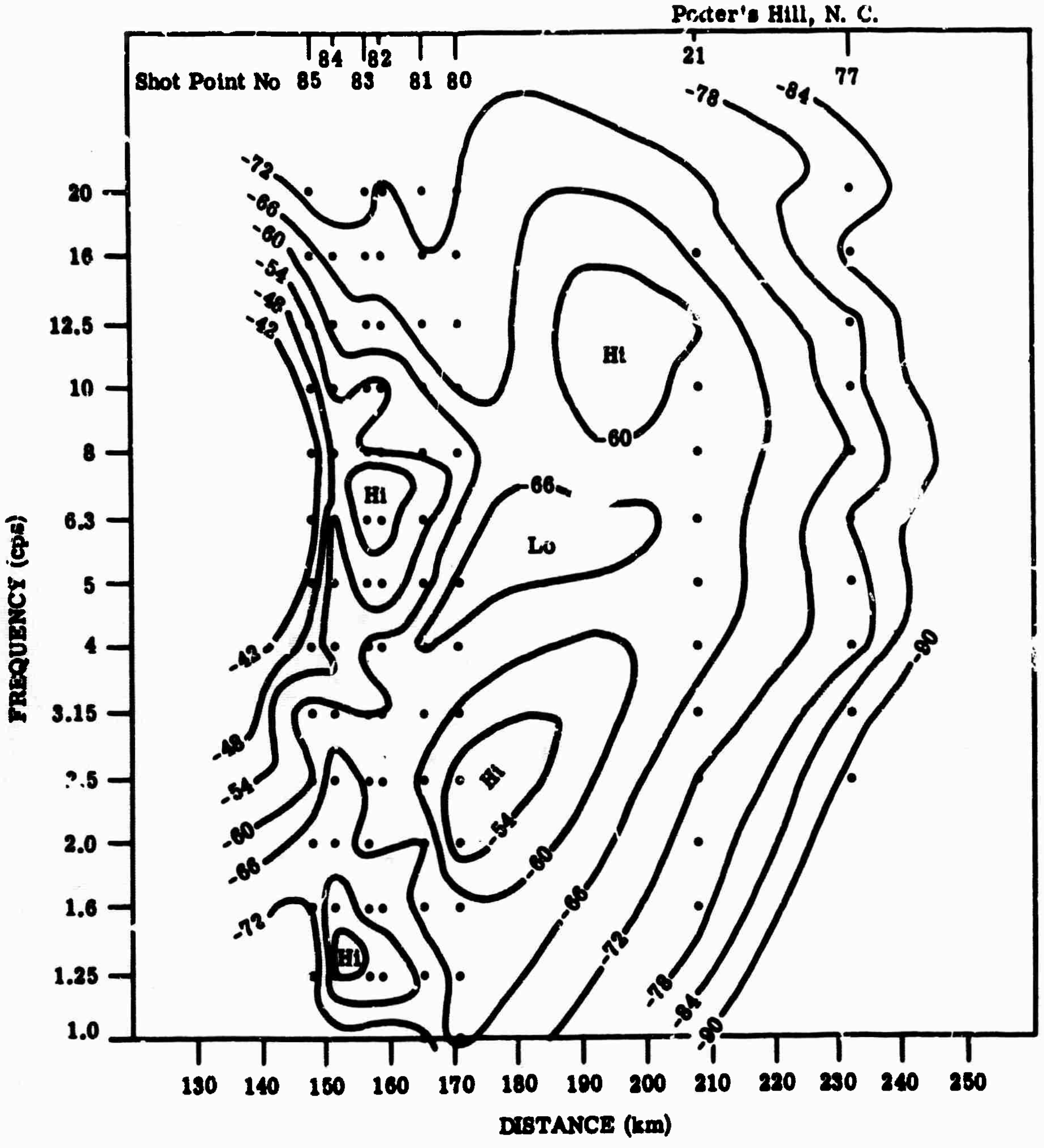

FIGURE 26. ISO-PARTICLE-VELOCITY MAP OF FIRST P-WAVE ARRIVAL FOR 1962 NORTH CAROLINA SHOTS RECORDED AT POTTER'S HILL, N. C.

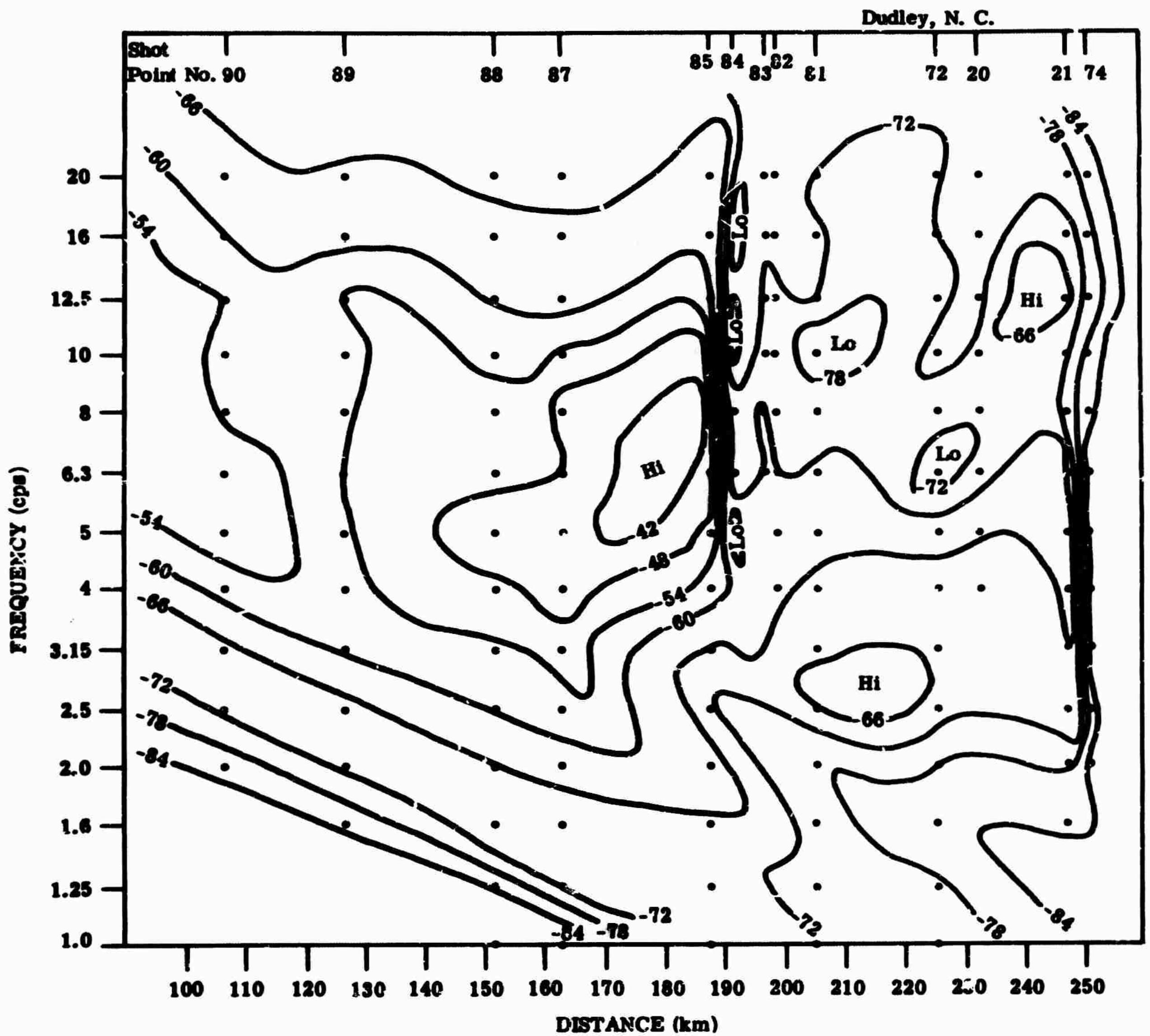

FIGURE 27. ISO-PARTICLE-VELOCITY MAP OF FIRST P-WAVE ARRIVAL FOR 1962 NORTH CAROLINA SHOTS RECORDED AT DUDLEY, N. C.

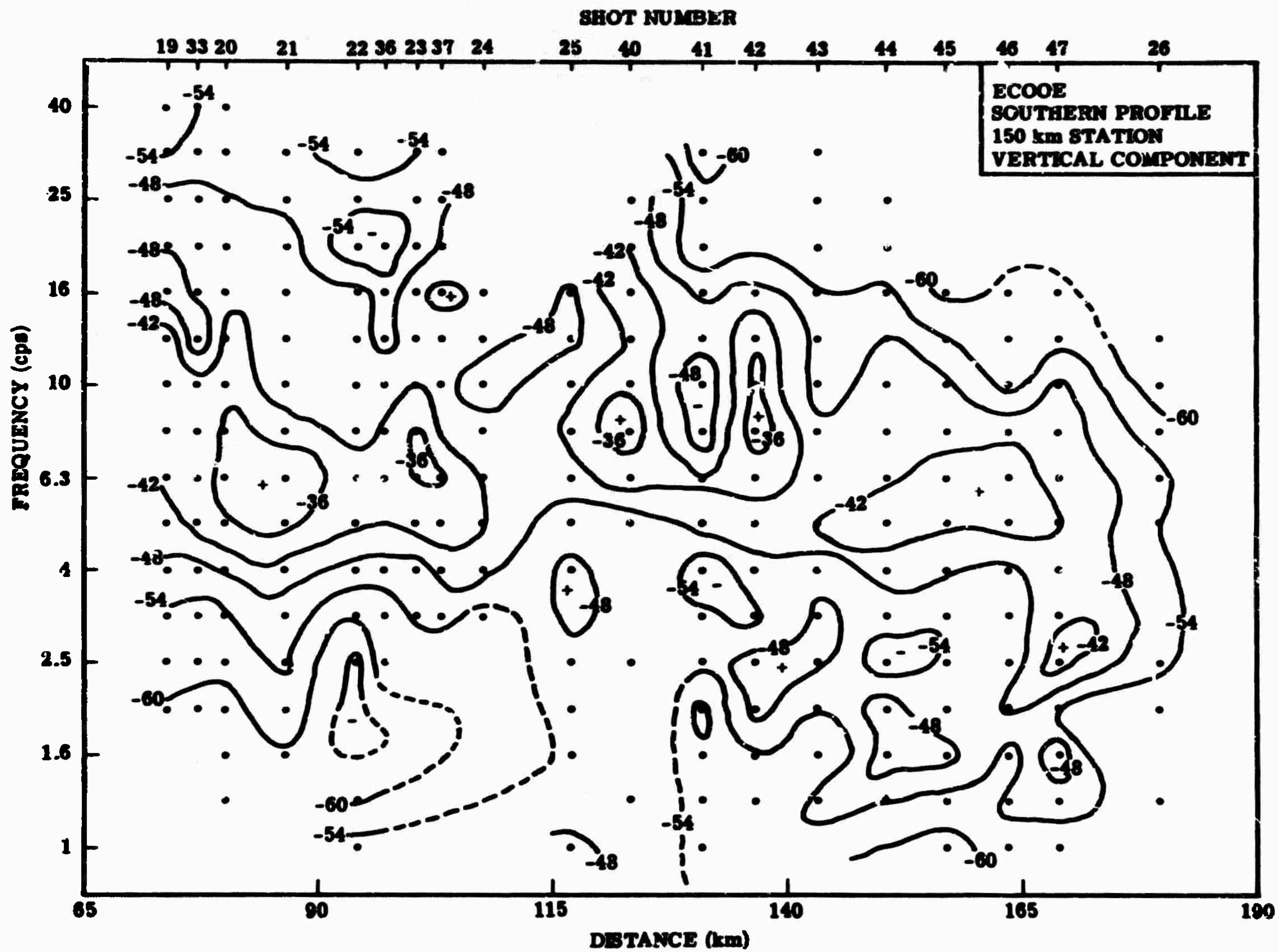

FIGURE 28. ISO-PARTICLE-VELOCITY MAP OF FIRST P-WAVE ARRIVAL FOR ECOOE SOUTHERN PROFILE SHOTS RECORDED AT POTTER'S HILL, N. C.

these data fall on the 6.0 branch of the travel time curve. There is not as much relief or change in signal level here as compared to the Early Rise or 1963 Lake Superior data. One reason is that the bubble pulse frequency is below 1.5 cps for all but the two farthest shots and the fundamental organ pipe frequency is above 10 cps for all but the last three shots. The particle velocity highs for the last three shots correlate with the bubble pulse frequency and the fundamental mode organ pipe frequency. All of these maps, without exception, displayed repeatable anomalies that correlated with specific shot points.

Residual iso-particle-velocity maps have been prepared to help isolate these anomalies and to determine the best values for the absorption and geometrical spreading losses. The source spectrum at a standard reference distance was computed by extrapolating the spectral data for the initial P arrival computed at each station to a distance of 1 km with assumed values of α and n. The average value of the particle velocity amplitude was then computed. With this as the source spectrum the predicted spectral curve for each shot was determined by means of equation 23 with the assumed values of α and n. The difference between the observed and the predicted signal can then be plotted and contoured similarly to the iso-particle-velocity maps. The residual maps having the smallest overall values probably represent the most appropriate values of α and n for the region. Examples of these residual iso-particle-velocity maps may be found in the paper by Willis and DeNoyer [1966]. The correlation

between the remaining anomalies, the bubble pulse frequency, and the constructive and destructive interferences caused by reverberations of the sound waves between lake bottom and the water's surface is quite plain.

In order to investigate in more detail the attenuation of the seismic waves using the type of data discussed above an equation of the type

$$A_{ij} = A_{oj}\, X_i^{-n}\, e^{-\alpha f_j X_i} \qquad (25)$$

was employed. It accounts for geometrical spreading (n) and for the absorption (α) of the seismic waves. A_{ij} represents the particle velocity amplitude for the i^{th} shot and the j^{th} frequency; A_{oj} represents a constant assumed to be independent of distance but dependent on the j^{th} frequency; X_i represents the distance from the seismometer to the i^{th} shot; and, f_j is the j^{th} frequency.

Initially we have an equation

$$A_{ij} = A_{ij}^{o}\, X_i^{-n}\, e^{-\alpha f_j X_i} \qquad \begin{array}{l} i = 1,2,\ldots, M \\ j = 1,2,\ldots, N \end{array} \qquad (26)$$

which can be solved for A_{ij}^{o}.

$$A_{ij}^{o} = A_{ij}\, X_i^{n}\, e^{\alpha f_j X_j} \qquad (27)$$

If we let

$$A_{oj} = \frac{1}{M} \sum_{i=1}^{M} A_{ij}^{o} \qquad (28)$$

we have arrived at equation (25).

Hence, for each A_{ij} an A_{ij}^{o} was calculated so that a matrix was obtained whose rows $(1,2,\ldots,j,\ldots,N)$ represent the N different frequencies, and whose columns $(1,2,\ldots,i,\ldots,M)$ represent the M different shot distances (in increasing order). An average A_{oj} was found, for each frequency, across the distances by summing the A_{ij}^{o} for a constant j and dividing the sums by M, the number of shots. There are then N A_{oj}'s which are in a respect independent of distance. This represents an average spectrum $(A_{o\hat{j}})$.

Using equation (25) and the computed A_{oj}, new A_{ij}'s designated A_{ij}') are computed, which in general will be different from the original A_{ij}'s.

The A_{ij}'s and A_{ij}''s are converted from particle-velocity amplitudes to the equivalent amplitudes in decibels by using the following relationship:

$$A_{ij}\ (\text{in/sec}) = 20 \log_{10} \frac{A_{ij}\ (\text{db})}{0.01} \tag{29}$$

A matrix of residuals (S_{ij}) are formed where

$$S_{ij} = A_{ij} - A_{ij}' \tag{30}$$

The S_{ij}'s are plotted on a graph where the abscissa corresponds to the distance and the ordinate is the frequency.

The summation of the residuals vs the absorption coefficient for values of n, the geometrical spreading factor, ranging from 0 to 3 are shown in Figure 29 for the Lake Superior 1963 South Line. The best fit for n and α would

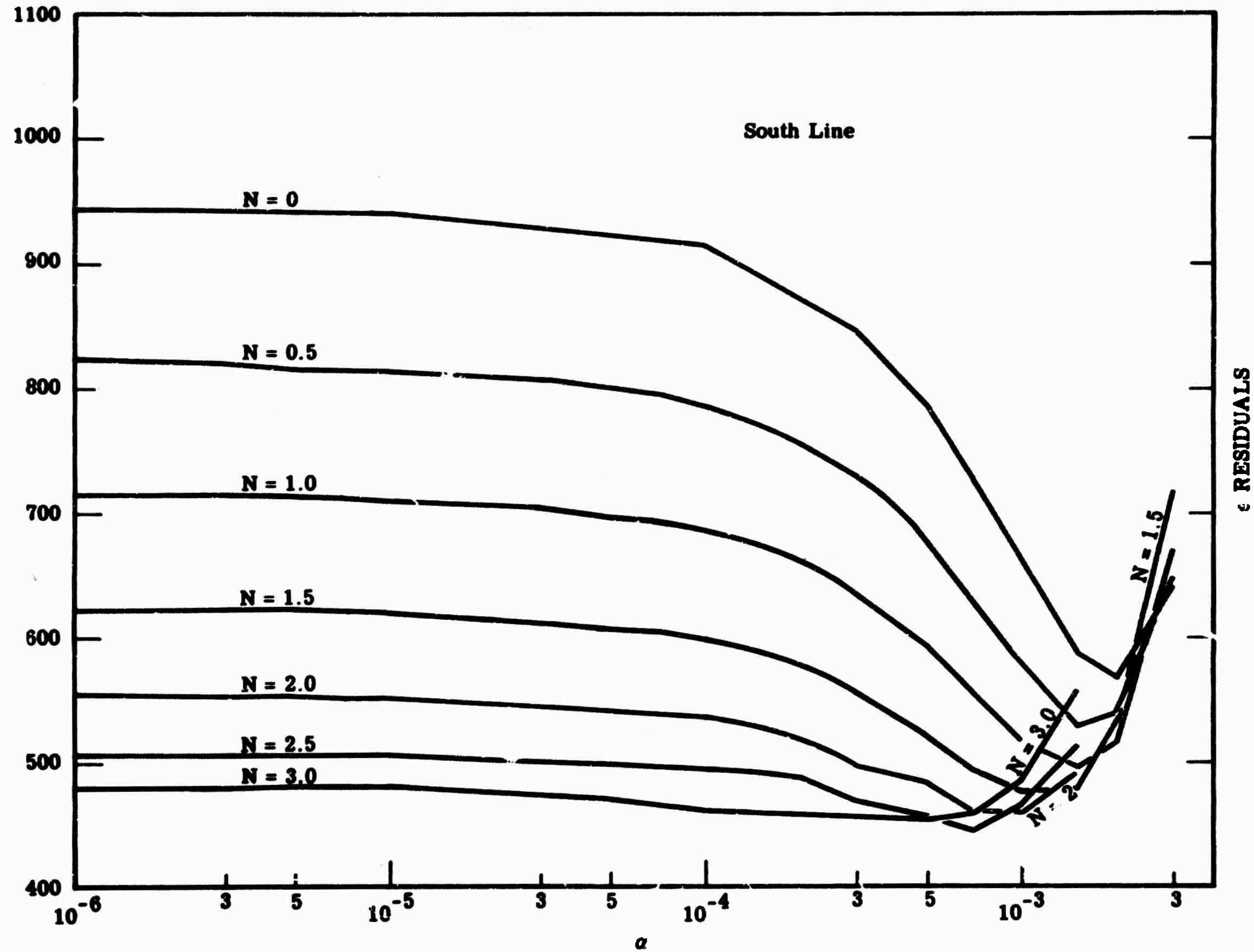

FIGURE 29. SUMMATION OF RESIDUALS vs α FOR 1963 LAKE SUPERIOR SOUTH LINE

correspond to the residual minimum. For this profile a value of $n = 2.5$ and $\alpha = 7 \times 10^{-4}$ yields the smallest residual. A value of $n = 2.0$ would correspond closely to the geometrical spreading due to head wave propagation. For $n = 2.0$, $\alpha = 1 \times 10^{-3}$ gives the best fit. This would correspond to a Q of approximately 490.

Q refers to the seismic anelasticity of a medium and can be defined as $\frac{2\pi E}{\Delta E}$ where E is the energy in the seismic wave. An equivalent expression given by Haskell [1957] defines Q^{-1} as $2\delta \times$ velocity which in the notation of this paper would be $\frac{\alpha \times \text{velocity}}{\pi}$ The spectrum of a seismic signal is a function of the source, the seismograph, and the transmission path. If the source function is known or is a constant and the response of the seismograph is known any systematic variations in the seismic signal can be correlated with transmission path affects. As mentioned earlier these affects are due to geometrical spreading and to anelastic absorption or scattering. The techniques described above were an attempt to determine the best fit for the geometrical spreading factor and the absorption. Depending on wave type, the geometrical loss correction factor should be easily removed from the attenuation data ($n = 0.5$ for surface waves, $n = 1$ for body waves and $n \approx 2$ for head waves). However, it was found that for first compressional wave arrivals, values of n other than 1 or 2 yielded smaller residuals in many instances. This variability in n strongly influences the values one would obtain for the

absorption coefficient. The Early Rise shots were the only ones where the source spectrum remained constant. Hence, caution should be used in comparing the values of Q from the different profiles.

The attenuation data for the 1963 Lake Superior South Line are somewhat anomalous compared to the other profiles described in this report. However, if only the data from the predominant frequencies are used (4 to 8 cps) the disagreement is reduced significantly. At 4 cps the best fit for this profile is given by a value of $n = 1.0$, $\alpha = 4 \times 10^{-3}$ ($Q = 123$), and $n = 1.5$, $\alpha = 3 \times 10^{-3}$ ($Q = 164$). On the 1963 Lake Superior east line and Knife River profiles we have values of $n = 1.5$, $\alpha = 3 \times 10^{-3}$ ($Q = 158$) and $n = 1.5$, $\alpha = 5 \times 10^{-4}$ ($Q = 894$) respectively. These minima agree closely with the minima obtained using all of the frequency data. It would thus appear that local seismic background noise adversely affected portions of the spectrum where the signal-to-noise level was low on the south line introducing a bias into the attenuation studies.

The Early Rise data were analyzed in a similar manner. Spectral analyses were limited to recordings made at distances less than 1068 km due to the poor signal-to-noise levels at greater distances. The vertical component of the first compressional wave arrival over this distance range had a minimum residual at $n = 1$ and $\alpha = 2 \times 10^{-4}$ ($Q = 1895$). For the second branch of the travel-time curve corresponding to the 7.7 km/sec layer the minimum residual occurred at a

value of $n = 1.5$ and $\alpha = 2 \times 10^{-3}$ ($Q = 204$).

Figure 30 shows the sum of the residuals versus α for n ranging from 0.5 to 3 for the ECOOE southern profile 150 km site. The best fit is for $n = 0.5$ and $\alpha = 6 \times 10^{-4}$ ($Q = 873$). For $n = 1.0$, a value of $\alpha = 3 \times 10^{-4}$ yields the smallest residuals. It can be seen in this figure that values of N greater than 1.5 can be discarded. However, even though the value of $n = 0.5$ gives a smaller residual than $n = 1.0$, the difference is not large so there is some uncertainty which value of n is most appropriate.

Similar graphs were prepared for the other recording stations of the ECOOE series. The average values of α for the minimum residuals are summarized in Table III.

Table III. Summary of ECOOE Attenuation Data

P	$n = 1/2$	$\alpha = 4 \times 10^{-4}$	vertical
		$\alpha = 3 \times 10^{-4}$	longitudinal
P_{Max}	$n = 1/2$	$\alpha = 5 \times 10^{-4}$	vertical
		$\alpha = 5 \times 10^{-4}$	longitudinal
		$\alpha = 8 \times 10^{-4}$	transverse
Maximum Surface Wave	$n = 1/2$	$\alpha = 8 \times 10^{-4}$	vertical
		$\alpha = 7 \times 10^{-4}$	longitudinal
		$\alpha = 1 \times 10^{-3}$	transverse

The Lake Superior data showed a degree of consistency with minimum residuals corresponding to values of $n = 1.5$ and Q varying from 158, 164, 204, and 894 for crustal paths. The latter (Knife River profile) also included waves which penetrated the upper mantle. The Early Rise profile which included long transmission paths through the mantle yielded

a value of Q equal to 1895. The east coast data showed larger crustal Q values and smaller upper mantle Q values than the Lake Superior data. Intuitively one would anticipate more reliable values of Q for the Lake Superior data since the east coast data are influenced by the Appalachian root system and the transition from oceanic to continental crusts and mantles.

The range in values of Q reported here is not inconsistent with published values. Anderson, et al [1965] estimates that the upper mantle Q is approximately 1500 for compressional waves. Sumner [1967] reports values of Q greater than 1000 for compressional waves in the Andes. O'Brien [1967] and Mueller [1968] give values of crustal Q's of 830 ±30 and 670 ±100 respectively. Teng [1968] reports values of Q of 450 in the upper mantle with a low Q zone of 60 at a depth of 50 km below the upper mantle boundary. In the western United States Long [1968] has found values of Q at distances between 100 and 600 km of 116 for the SHOAL event and 169 for the GNOME event. Other investigators (Anderson and O'Connell [1967], Anderson [1967], Anderson [1966], and Walsh [1968]) have studied shear and surface wave Q values in the mantle as a function of depth and temperature. Their studies, however, are restricted primarily to long period data. Average values of Q for surface waves in the upper 400 km of the mantle range between 88 and 135. Shear Q values average about 200 for the upper 600 km of the mantle and about 600 for the entire mantle.

Pakiser [1965] has shown that the Rocky Mountains divides the crust and upper mantle of the United States into two superprovinces. The eastern United States has a thicker average crust with larger crustal and P_n velocities. This region also has larger magnetic anomalies indicating an abundance of magnetic materials. The higher velocities in the lower crust could be explained by lateral variations of eclogite to amphibolite. The average upper silicic crust is about the same thickness (19 km) in the eastern and western United States while the average lower mafic crust is thicker in the eastern United States.

The higher crustal and mantle P velocities of the eastern United States correlate with the larger values of Q found in this same area. The smaller upper mantle Q values observed in the western United States are probably due to the low velocity layer in the upper mantle although it may also be due in part to lateral variations of eclogite-peridotite in the upper mantle. Since high temperatures cause the attenuation of seismic waves to increase rapidly (Anderson [1967]), the higher heat flow measurements in the western United States may also have some correlation to the lower Q values.

In Figure 31 the summation of residuals versus n is presented where the absorption coefficient, α, was set equal to zero. All of the attenuation of the seismic waves thus is accounted for by the geometrical spreading factor. The minima vary from n = 1 (ECOOE) to n = 3.5 (Lake Superior

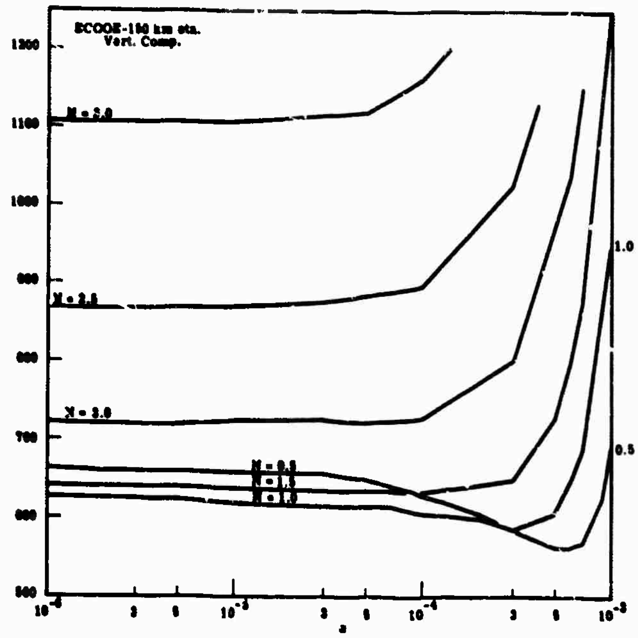

FIGURE 30. SUMMATION OF RESIDUALS vs α FOR ECOOE SOUTHERN PROFILE 150 km SITE.

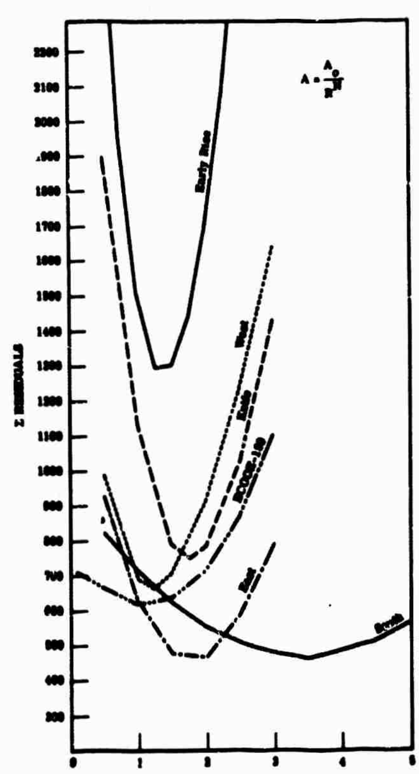

FIGURE 31. SUMMATION OF RESIDUALS vs n.

South Line) and average approximately 1.5. Hence, geometrical spreading by head waves (n = 2.0) would cause too great a loss in seismic energy to account for the observed loss at most of these sites. The head wave amplitudes were either too small to be observed or perhaps the upper mantle boundary is too uneven to support their transmission.

Additional Studies

The investigation of the attenuation of seismic waves discussed in this report did not take into consideration the effect of the changing angle of incidence of the seismic wave as a function of distance from the source. Three component data were obtained for all of the shots reported here. Particle motion diagrams have been prepared for the first compressional wave arrivals. A linear least-square regression analysis technique was investigated for estimating the parameters affecting the attenuation of seismic waves. A negative absorption coefficient for the longitudinal component was disclosed by the least square approach which implies that the vertical and longitudinal component particle velocities should be combined vectorially and that the transformed data should yield better estimates of the propagation losses.

The problem of shot point and recording station anomalies should be investigated further to determine the best approach for removing these affects in computing the absorption coefficients. The former anomalies are not too critical where the same shot point is used (e.g., the Early Rise shots). However, where moving shot points and fixed recording stations are used (e.g., 1963 Lake Superior Experiment and ECOOE), the shot point anomalies can frequently mask or seriously confuse the normal attenuation losses.

Another topic that needs further investigation is the particularly good coupling of seismic energy found in the

Lake Superior area. Current studies are being conducted that may give some insight into this subject. A series of recordings were made in Lake Superior on the lake bottom near shot point 51 during the summer months of 1967. These recordings were made with modified ocean bottom seismographs in an effort to determine if there was any reciprocity in the propagation of seismic waves from teleseismic events. These recordings are currently being analyzed.

Conclusions

1. Seismic refraction studies disclosed that the earth's crust varies in thickness from 28.1 km at the continental margin in North Carolina to 50.4 km near shot point 51 in Lake Superior. A single-layer crust with a compressional velocity of 6.0 km/sec was found in North Carolina while a two-layered crust with compressional velocities of 6.2 and 7.7 km/sec was found in the Lake Superior area. Larger travel time residuals for recording stations located i the Appalachian Mountains are indicative of a mountain root system similar to the Rocky Mountains. No indication of a low velocity layer was found in the upper mantle in the eastern United States that would correlate with such a feature reported in the western United States.

2. The Lake Superior area was found to be more efficient in the coupling of seismic energy for underwater shots than the area in the Atlantic Ocean off the coast of North Carolina. Signals ranged up to a factor of eight larger.

3. A new method of displaying seismic attenuation and spectral data was developed that permits the presentation of the data in three-dimensional space—distance, frequency, and ground particle velocity amplitude. This technique allows the presentation of large quantities of data in a concise manner that assists in the interpretation of the data.

4. Meaningful and consistent anomalies in the spectrum of seismic waves generated by underwater shots were found that could be correlated with variations in the source region and in propagation effects.

5. The maximum amplitude of the first compressional wave arrivals attenuate like the inverse 3/2 power of the distance out to approximately 1100 km from the Lake Superior area. Conversely an inverse cube law applies to shots fired in the Atlantic.

6. The detailed spectral studies of the Lake Superior shots disclosed that the crustal attenuation data could be fit using a geometrical spreading factor of $n = 3/2$ and $\alpha = 2\text{-}3 \times 10^{-3}$. Mantle compressional waves fit a geometrical spreading function of $n = 1$ and $\alpha = 2 \times 10^{-4}$. These give average upper crustal Q values of approximately 160, a lower crustal Q of approximately 200 and an upper mantle Q of approximately 1900. The Atlantic Ocean shots had values of $n = 1.2$ and $\alpha = 6 \quad 10^{-4}$ to 4×10^{-4}. The corresponding values of Q would equal 873 and 960. The former includes transmission paths predominantly within the crust while the

latter has transmission paths predominantly in the mantle. Hence, the east coast data shows the same order of magnitude values for Q in the crust and upper mantle.

7. Attenuation losses in the propagation of seismic waves did not support the theory of head wave propagation which would require a decrease in amplitude approximately proportional to the inverse square of the distance.

8. The variation in values for the geometrical spreading factor and absorption coefficient caused by the broad minimum residuals indicate that scattering effects, station anomalies and source variations play an important role in the apparent attenuation of seismic waves and that methods of determining the influence of these factors will have to be developed before more accurate determinations of Q can be made over the frequency range of .5 to 25 cps.

APPENDIX A

Underwater Explosion Experiments

The series of underwater shot programs listed in Table I were participated in by a number of organizations that banded together in the loosely knit group called the North American Explosion Seismology Group. Participation of the members varied with each series of shots. The participants are listed below.

Boston College
Carnegie Institution of Washington
Dominion Observatory, Canada
Georgia Institute of Technology
Geotech Division, Teledyne Industries
Pennsylvania State University
Princeton University
Stanford Research Institute
Southwest Center for Advanced Studies
University of Alberta
University of Manitoba
University of Michigan
University of Minnesota
University of Toronto
University of Western Ontario
University of Wisconsin
U. S. Geological Survey
Vela Seismological Center, AFTAC

The 1962 North Carolina Experiment was conducted during the months of July and August. The program consisted of a series of 39 one-ton shots fired on the ocean bottom along a profile normal to the coast near Cape Fear, North Carolina.

The shots were fired on and off the continental shelf. Details of this experiment may be found in a paper by Meyer, et al. [1965].

The 1963 Lake Superior Experiment consisted of a series of 78 one-ton shots fired along a line extending the length of Lake Superior. All shots were fired on the lake bottom. Details of this experiment have been described by Steinhart [1964].

The October 1964 Lake Superior Experiment consisted of a series of 1, 2, 3, 6, and 10 tons of Nitramon WW-EL fired on the bottom at shot points 45-49 and 73 of the 1963 series. A total of 10 shots were fired. One shot of each size was fired at shot point 45.

The East Coast On-Shore Off-Shore Experiment (ECOOE) consisted of a series of one-, five-, and ten-ton shots fired along four lines, two off the coast of North Carolina and two off the coast of Virginia. Details of this experiment may be found in a report by Hales, et al [1967].

The last large scale cooperative experiment consisted of the 38 five-ton shots of the EARLY RISE series. These were fired in Lake Superior on the lake bottom near shot point 51 of the 1963 series. Details of this program are in a report by Warren, et al [1967].

The CHASE (<u>c</u>ut <u>h</u>oles <u>a</u>nd <u>s</u>ink <u>e</u>m) series consisted of three shiploads of surplus explosives that were disposed of by the U. S. Navy. The explosive-laden ships, primed with pressure detonators, were scuttled in deep water off the

Atlantic Coast. Pertinent details may be found in Table IV.

Table IV. CHASE Series*

Event	Date	Time	Location	Yield (tons)	Magnitude	Depth of Shot (ft)	Depth of Water (ft)
III	7/15/65	1416:08Z	37°11'48"N 74°21'06"W	700	4.63 ±0.53	900	5000 (est.)
IV	9/16/65	1951:10Z	37°11'34"N 74°26'34"W	300	4.73 ±0.39	900	5100
VII	7/29/66	0436:26Z	36°29'56"N 74°13'42"W	400	4.62 ±0.56	3000	7500

*Data from LRSM Reports

APPENDIX B
Instrumentation Characteristics

Matched three component seismometers of the Hall-Sears Model HS-10 2 cycle type were used for most of the data presented in this report. High impedance coils (200,000 ohms) were used for high gain purposes. The transistorized seismic amplifiers used were constructed by the Geophysics Laborabory and had a maximum gain of 114 db. The bandwidth of these instruments was essentially flat from 0.2 to 500 cps and had an input noise with a 5 megohm source of less than 1 μV. A variety of instrumentation type FM tape recorders were used that had an average signal-to-noise ratio of 40 db. The standard practice used in the field was to record the output of each seismometer at two or three levels separated by 12 or 18 db to provide wider dynamic range. A more detailed description of this equipment may be found in Willis and Jackson [1966], Willis and Johnson [1959], and Willis [1966].

APPENDIX C

Spectral Analyses Techniques

The frequency analyses data presented in this paper were obtained by bandpass RLC filtering. Different spectral analyses techniques were investigated. These include

- Bandpass RLC filtering
- Fourier transform analysis employing an electronic analog computer
- Digital analyses using an IBM 7090 employing Fourier transform methods
- Optical diffraction of variable-density film records.

A comparison of the first three techniques may be found in a paper by McIvor [1964]. The optical analysis technique is described in papers by Jackson [1964] and Jackson [1965]. The results obtained from these different approaches are in good agreement if the data are normalized to the same bandwidth. The digital, analog and optical computer technique provide finer frequency resolution than the bandpass filtering method. However, for low frequency seismic signals (below 10 cps) this fine frequency resolution usually is not required.

For the data reported in this report the spectral information was performed by playing back the demodulated FM magnetic tape signals through a series of electronic polyfilters, each filter of which was spaced at one-third-octave intervals. These filters are composed of three transistorized cascaded RC networks with active feedback elements. The outputs are passed through DC line amplifiers to bifilar

galvanometers with a natural frequency of 500 cps. The galvanometer deflections are recorded on a conventional photographic recorder. The time window of these filters is approximately Gaussian. The use of the DC line amplifier permits amplification of the signal thus facilitating the interpreters measurement of the trace deflection (providing of course, that the signal is not in the seismic background noise or the noise of the recording and playback systems). This permits more accurate measurements and in many cases allows the interpreter to measure signals that would otherwise be omitted because of low signal levels. One can thus examine teleseismic signals that contain energy well below the natural frequency of the seismometer. This is illustrated in Table V.

Shown in Table V are the signal levels (*) for the first compressional wave arrival of a teleseismic event as measured on each one-third-octave seismogram over the frequency range from 0.315 to 2.5 cps. At 0.4 cps the signal level read on the third-octave seismogram was -10 db (peak-to-peak tract deflection of 0.632 inches). If no playback gain were used, this signal level would have been -30 db (peak-to-peak trace deflection of 0.063 inches). The signal level was 10 db (factor of 3.2) above the background noise. Hence, it is evident that it is much more accurate to read a displacement of 0.632 inches with an average background noise of 0.2 inches than 0.063 inches with a background of 0.02 inches. A signal of this amplitude would often be overlooked

especially if one were measuring the signal level on the next lower frequency seismogram (0.315 cps).

Column (a) shows the corrected seismometer signal output which includes corrections for the record gain used in the field recording and the playback gain used in the laboratory. Column (b) takes into account corrections for the seismic preamplifier response and the seismometer response. The signal levels shown here would be for an amplifier and a seismometer whose response would be flat between 0.315 and 2.5 cps. Column (a) shows a peak signal level at 0.8 cps. At 0.5 cps the signal level is 11 db lower. However, from column (b) it can be seen that after corrections for instrumentation response are made, the difference in signal level between 0.8 and 0.5 cps is only 1 db. Hence, the true energy has a broader spectral width than is apparent on a broadband seismogram. The peak signal is below the seismometer's resonant frequency and hence any variation in frequency of this signal from one recording site to the next would be magnified or exaggerated on a broadband or regular seismogram (for example, see Willis 1965). This variation would not be as evident or might be completely masked on the latter type record.

Table V. Spectral Corrections

f (cps)	Signal* Level (db)	Record Gain (db)	Playback Gain (db)	Corrected Signal Levels: a — From Seismometer (db)	b — Corrected for Amplifier and Seismometer Response** (db)	c — Normalized to 1 cps Bandwidth (db)
2.5	-4.5	70	20	-94.5	-76.5	-77
2.0	-1	70	20	-91	-73	-73
1.6	-0.5	70	16	-86.5	-68	-67.5
1.25	0	70	16	-86	-67	-65.5
1.0	0	70	14	-84	-63.5	-60
0.8	-0.5	70	6	-76.5	-52	-48
0.63	+0.5	70	12	-81.5	-51	-47.5
0.5	+0.5	70	18	-87.5	-51	-46.5
0.4	-10	70	20	-100	-56	-49
0.315	-19	70	20	-109	-57	-50

*A peak-to-peak trace deflection of 2 inches = 1 V_{rms} which is referred to as a standard reference (0 db).

**Seismometer response corrected to a flat output of 1 V_{rms} at a constant velocity of 1 x 10^{-2} inches per second.

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13. ABSTRACT

This paper describes the travel-time anomalies and attenuation losses of seismic compressional waves generated by a series of underwater explosions in the eastern United States. The efficient tamping of the shots fired in water provided a seismic source that could be detected at much larger ranges than could be accomplished by equivalent sized shots fired underground.

A number of mobile field recording stations equipped with three-component matched short period seismometers and magnetic tape recorders were used to record 273 individual shots. A total of 1,295 recordings were obtained along a reversed profile extending from International Falls, Minnesota to the Atlantic Coast. The underwater shots were fired in Lake Superior and the Atlantic Ocean. Precise travel times were obtained by recording radio time signals at all of the recording stations.

An analysis of the travel-times of the seismic waves disclosed that the earth's crust varies in thickness from 28.1 km near the Atlantic Coast in North Carolina to 50.4 km near the Keweenaw Peninsula in upper Michigan. The crust in North Carolina was found to be comprised of a single layer with a compressional wave velocity of 6.0 km/sec. A two-layer crust with compressional wave velocities of 6.2 and 7.7 km/sec was disclosed in upper Michigan. Travel time residuals across the Appalachian Mts. would indicate a mountain root system similar to that found under the Rocky Mts.

The Lake Superior area was found to be more efficient in the coupling of energy into seismic waves by the underwater shots than in the Atlantic Ocean. This is believed due mainly to the bubble pulse phenomenon and the reinforcement caused by reverberation of sound waves between the lake bottom and the water's surface.

A new technique of displaying seismic attenuation data in three dimensional space was developed that permits presentation of large quantities of data in a concise manner, hence aiding in the interpretation of the data. Consistent anomalies in the spectrum of the seismic waves were found that could be correlated with various parameters in the source region and with propagation effects.

The detailed spectral studies that were made as a function of distance from the source disclosed that the attenuation of the seismic waves could be expressed by the equation

$$A = A_0 X^{-n} e^{-\alpha f X}$$

where A = amplitude, X = distance, n = geometrical spreading factor, α = absorption coefficient and f = frequency. The crustal attenuation data in the Lake Superior area were found to have a value of $n = 1.5$ and $\alpha = 2\text{-}3 \times 10^{-3}$. Upper mantle compressional wave attenuation data gave a value of $n = 1$ and $\alpha = 2 \times 10^{-4}$.

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Data analysis						
Field measurements						
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Travel times						
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